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ANOMALOUS WARMING OF THE STRATOSPHERE OVER NORTH AMERICA IN EARLY 1957

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ABSTRACT

The anomalous stratospheric warming of January and February 1957 is studied in detail by means of constant pressure charts, time sections, and cross sections. The warming occurred as the meandering Arctic stratospheric jet stream of wave number two developed into a pair of vortices extending to above the 10-mb. surface. It is concluded that development of stratospheric waves in the Northern Hemisphere is facilitated by autumnal growth of a forced perturbation locked in place by a warm ridge over the Aleutian area. The baroclinicity and rate of development were in agreement with Fleagle's criteria for the growth of disturbances. Interaction between the tropospheric jet stream and Arctic stratospheric jet stream during the period of development is believed responsible for the great intensity of the 1957 warming.

1. INTRODUCTION

For the meteorologist familiar with the synoptic charts of the lower levels, the annual course of events in the upper stratospheric layers,[1] as represented by constant pressure analysis of surfaces between 50 and 10 mb., is a fascinating one to follow. During the summer months, when temperature increases from equator to poles, east winds that increase with height blow in almost complete isolation from activity in the lower stratosphere and troposphere. In September and October [27], winds with westerly components appear first in high latitudes and then progressively southward into the subtropics. The troughs and ridges of the upper stratospheric westerlies generally conform with the major troughs and ridges of the troposphere and are influenced little by minor tropospheric waves. However, during the coldest part of the year, traces of strongly developing systems may appear at the 50-mb., or rarely at the 25-mb., surface, although with marked westward tilt and with amplitude greatly reduced particularly above the southern portion of the tropospheric disturbances. Many of the relationships and differences between stratospheric and tropospheric circulations were first described by Austin and Krawitz [1].

Within the Arctic Circle, the gradual lowering of temperature during the first months of the long polar night is accompanied by formation and intensification of the Arctic stratospheric jet stream in the baroclinic zone between the cold Arctic core and the warm zone that almost completely encircles the earth above the tropopause in higher middle latitudes. The speed of this jet stream quite regularly reaches 100 knots and occasionally exceeds 200 knots.

A large increase in amplitude of waves in the Arctic stratospheric jet stream and accompanying deterioration of its cold core have been observed to occur quite suddenly at some time during the first 12 weeks of the year, but there have been insufficient analyses of the 50-mb. and higher surfaces for the past years to reveal all of the necessary stages in the life history and other general characteristics of the Arctic stratospheric jet stream. The

[1] Herein, the stratosphere will be considered as the layer from the tropopause up to the stratopause. In middle latitudes this layer is nearly isothermal. The stratosphere is divided into the upper stratosphere, approximately from 35 km. down to 20 km., and the lower stratosphere, from 20 km. down to the tropopause. In contrast to the upper stratosphere, the lower stratosphere exhibits strongly many of the minor trough and ridge features of the troposphere. In upper stratospheric troughs and ridges, contours and isotherms are generally in phase while in the lower stratosphere contours and isotherms are usually out of phase. Above the stratopause is the thermoincline of the ozonosphere or chemosphere in which temperature increases with height up to a maximum somewhere near the 50-km. level. The stratopause at low latitudes is indistinct since the tropical stratosphere is also a thermoincline, while at high latitudes, in winter, its nature is unknown.

most spectacular phenomenon associated with the amplification of waves in this jet stream has been anomalous warming of the upper stratosphere. Scherhag [17] reported an instance of this type of anomalous warming in 1952 and referred to it as "explosive" warming. Possibly because the large scale of the phenomenon in process was not fully appreciated, a solar origin for the warming was proposed by some writers. Following further documentation of warming epochs, Wexler [29, 31] found that little if any convincing evidence of a solar origin could be brought forth and that the observed warming could in each case be explained by subsidence and the adiabatic heating of air descending from potentially warmer layers. The possibility of a trigger action by solar activity still remains since the polar stratosphere seems to build up in the winter season to a condition of baroclinic instability and not all of the factors capable of releasing the instability and producing associated phenomena like anomalous warming are known.

The stratospheric warming of 1957 is perhaps the most spectacular of those reported on to date. It certainly lends itself to closer examination than those that occurred previously because of its initiation in middle latitudes over North America and subsequent passage over many excellent radiosonde stations and because of the exceptional magnitude of associated features such as wind speeds and temperature changes. Craig and Hering [2] presented selected charts from their series of daily 100-, 50-, and 25-mb. charts to show the changes that took place at one week intervals between January 23 and February 20, 1957. Godson and Lee [6] discussed developments in this same period and for analogous periods in the two previous winters by means of vertical cross sections and time-altitude charts. Even so, these complex motions of the atmosphere have not been completely explained.

Any attempt to analyze the circulation of the upper stratosphere during an entire month or season and to relate it to events in the lower layers extending down to sea level leads to a multiplicity of 2-dimensional diagrams describing sections through a 4-dimensional situation. In this paper, references to events in the lower troposphere have been held to a minimum as one means of reducing the complexity of the discussion. Many of the diagrams are composites constructed from data that have been interpolated, extrapolated, and averaged, in space or time, to minimize the effects of the sparsity and inaccuracy that are common to reports from very high altitudes. These factors were less of a problem in this 1957 case because of the large number of well-equipped stations in the region where the 1957 warming occurred and the greater heights that are progressively being attained by radiosondes.

2. SOME SOURCES OF ERROR

At this point a few remarks concerning the quality of radiosonde data are in order. These statements have been synthesized from the literature, from expert opinion, and from studies being completed by the author and his colleagues. Stratospheric temperatures are in most cases probably accurate to within 3° or 4° C. This accuracy with the U. S. Weather Bureau duct-type radiosonde in daylight flights to very great heights is obtained only by means of a radiation correction that tends to make this instrument compatible with those having a white external temperature element. Uncorrected daytime temperatures that may still be transmitted by some nations may involve very large errors at great heights. The residual diurnal variation of temperature after correction is still a burdensome complication in the analysis of time sections and other charts.

Another source of difficulty in stratospheric analysis is the pressure error. In the lower atmosphere, an error of a few millibars does not have much effect on the temperature at a level or on the computed heights of standard pressure surfaces and significant points such as fronts and the tropopause. However, at the 10-mb. surface each millibar of error causes a height error of about 650 meters in the placement of a temperature or significant point. There is evidence in the data that errors of 3 mb. are not uncommon and in a few soundings the errors appear to be much larger. The accuracy of wind calculations is affected by pressure error; the result is a wildly oscillating wind speed in cases where the pressure error is variable due to a sticking, intermittently operating baroswitch. The percentage error in calculated rate of ascent of radiosonde is nearly the same as the percentage error in pressure and for some conditions equals the percentage error in wind speed. Relatively little error in temperature and computed height is produced by the pressure error in layers with nearly isothermal lapse rate. Many of the problems that are caused by imperfect, incomplete data have been encountered and solved in summary fashion in the course of earlier stratospheric analysis projects [8, 12].

3. EVENTS LEADING UP TO THE "EXPLOSIVE" WARMING OF THE STRATOSPHERE

In January 1957, radiosonde data were being transmitted only up to the 25-mb. level on teletypewriter, and in fact at only a small percentage of stations was radiosonde equipment capable of regularly reaching that level. Thus, it was not until January 24, when large 25-mb. temperature rises were reported from Newfoundland, that the "explosive" warming of 1957 became detectable to those making use of current data received by teletypewriter. For this reason, and also because it was on that date that a well-formed, regularly-moving, warm center became established, January 24 is taken as the initial date of explosive warming. However, it will be shown below that marked warming had already been observed very high in the stratosphere at Salem, Oreg. at least 5 days earlier.

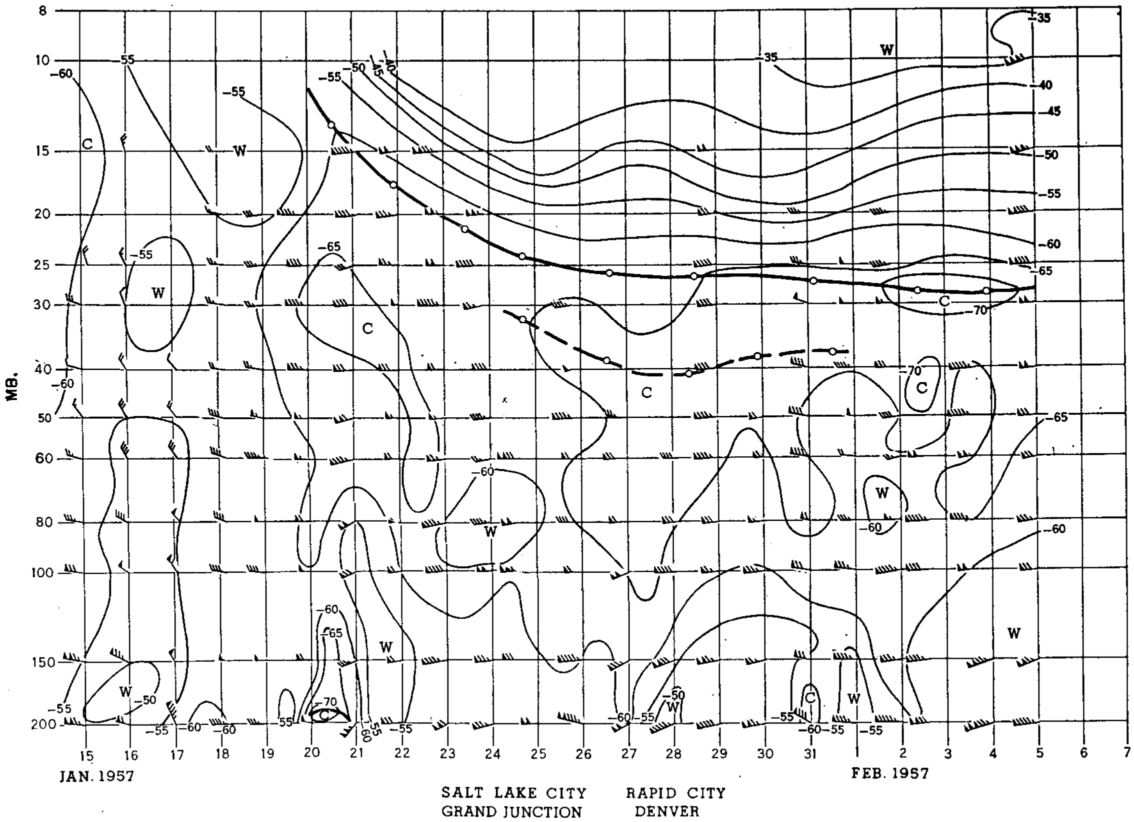

Figure 1.—Time section, January 15 to February 5, 1957, for Denver, Colo. supplemented by data for Salt Lake City, Utah, Grand Junction, Colo. and Rapid City, S. Dak. Isotherms are thin lines labelled in degrees Celsius; wind speeds are given by barbs for each 10 kt. and triangles for each 50 kt. on a shaft oriented from top of diagram for north winds. Medium line with superimposed circles indicates stratopause; line is broken to show alternative stratopause. Medium weight line without circles indicates tropopause. Vertical lines are located at 1500 GMT of indicated date. The analysis of this and other charts is carried across portions where interpolation seems justified but not into portions requiring extrapolation of data.

At the beginning of January 1957, low pressure in the upper stratosphere was centered near the North Pole, and a portion of the cold wintertime airmass of the Arctic stratosphere had penetrated southward in a trough extending from northwestern Greenland to Newfoundland and thence into the North Atlantic Ocean. At 25 mb. warm air and an anticyclone were centered over western North America. As the month progressed, cold air was advected farther into the western portion of the trough which gradually retrograded westward into central Canada. On January 12, a closed Low appeared in this trough just north of Hudson Bay. By January 18, cold air from the northwest was entering the middle stratosphere over the northwestern United States. Some of the warm air that had been associated with the anticyclone over western North America was moving rapidly eastward across the northeastern States, and the anticyclone itself had melted away into the subtropical Pacific Ocean area. Although normal contours and isotherms for the upper stratosphere have not been established, reference to daily and mean charts for and above the 50-mb. surface [1, 3, 7, 14, 23, 26] suggests that isotherms and contours at this time were very anomalous over the United States and Alaska. In comparison to the average, the temperature over parts of Alaska was 20° C. colder than usual in January. The cyclonic curvature, low temperatures, and strong westerly wind components in the northwestern United States also appear to have been anomalous.

The appearance of upper stratospheric warming on January 19 at Salem, Oreg., was first pointed out to the author in a private communication from R. A. Craig and W. S. Hering, who also described its rapid passage across the United States in the days that followed. At Salem, the reported 9-mb. (32-km.) temperature on January 19 was −48° C., an increase of 21° C. in 5 days, while at a lower altitude in the same observation a temperature of −66° C. was reported at 16 mb. (28 km.) in the southward-moving Arctic stratospheric airmass. Increasing stratospheric winds exceeded 100 kt.

Warming of the upper stratosphere and local acceleration of the wind frequently occur simultaneously with cooling of the lower stratosphere at the advancing edge of the Arctic stratospheric airmass. The lapse rate within this airmass is generally positive but very small, perhaps 1° C. per km. Above in the thermoincline there may be a temperature inversion of as much as 5° C. per km.

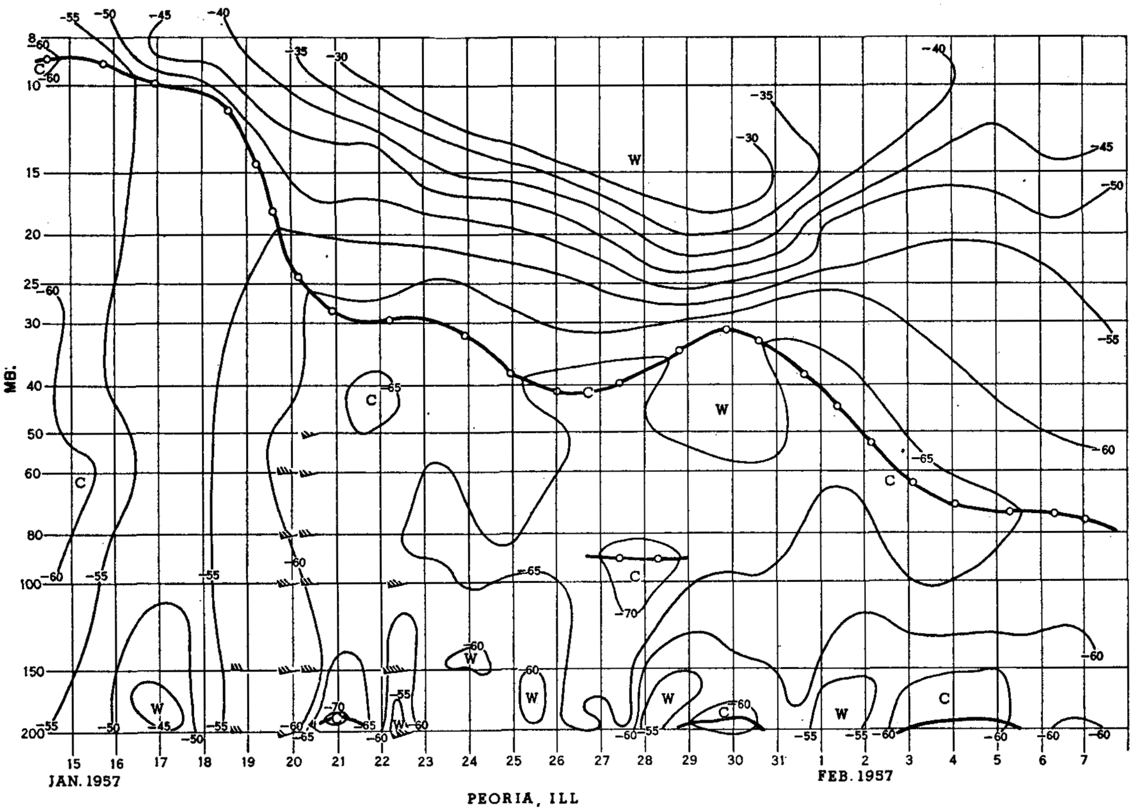

Figure 2.—Time section, January 15 to February 7, 1957, for Peoria, Ill.

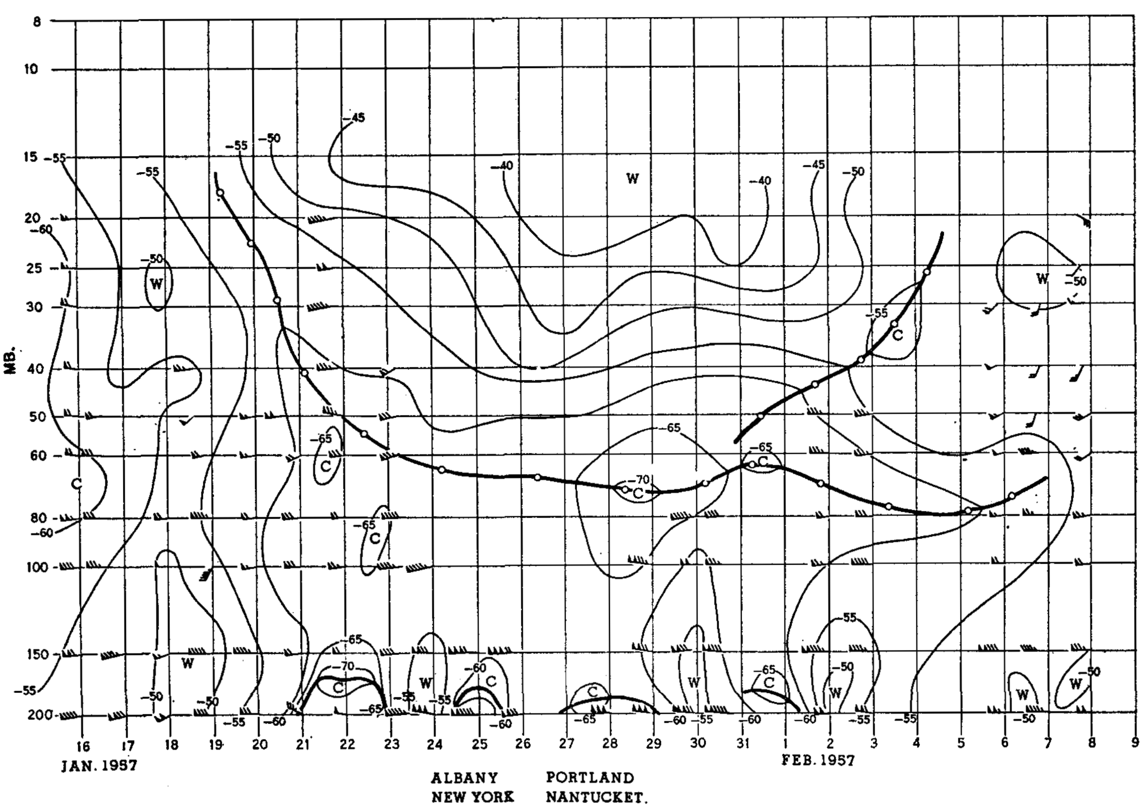

Figure 3.—Time section, January 16 to February 8, 1957, for Albany, N. Y., supplemented by data for Portland, Maine, New York City, and Nantucket, Mass.

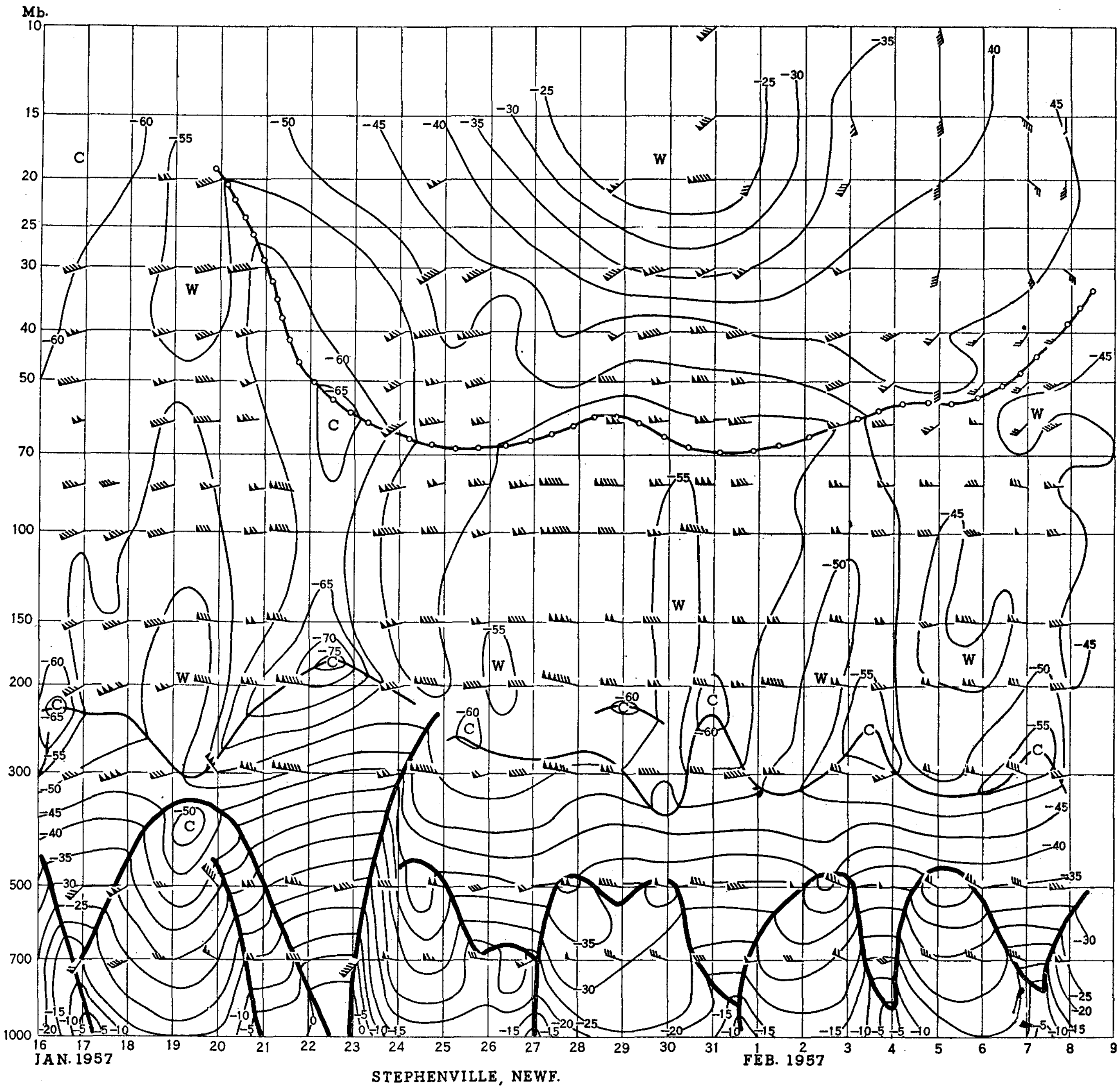

Figure 4.—Time section, January 16 to February 8, 1957, for Stephenville, Newfoundland. Analysis as in figure 1, but including frontal positions (heavy lines).

The change in lapse rate is often abrupt, forming a clearly marked boundary surface that will be referred to here as the stratopause. In this special case, the altitude of the stratopause may be much lower than usual and is analogous to the low tropopause found above cold tropospheric airmasses.

In figure 1, a composite time section for the east-central Rocky Mountain area, based primarily upon data from Denver, Colo., the stratosphere on January 19–20 cooled to below −65° C. Within 2 days, warming to −40° C. had occurred at 11 mb. and the wind speed at the high levels had increased to more than 100 kt., a high value not often observed at this altitude and latitude. The stratopause, after making its appearance below the 10-mb. surface, descended to the layer between 25 and 30 mb. by January 26. In contrast, the stratopause at Salem remained in the layer between 12 and 22 mb. from January 19 until the period February 2–6, when it descended to below the 50-mb. surface. Below the stratopause, temperatures in the Arctic stratospheric airmass at Denver

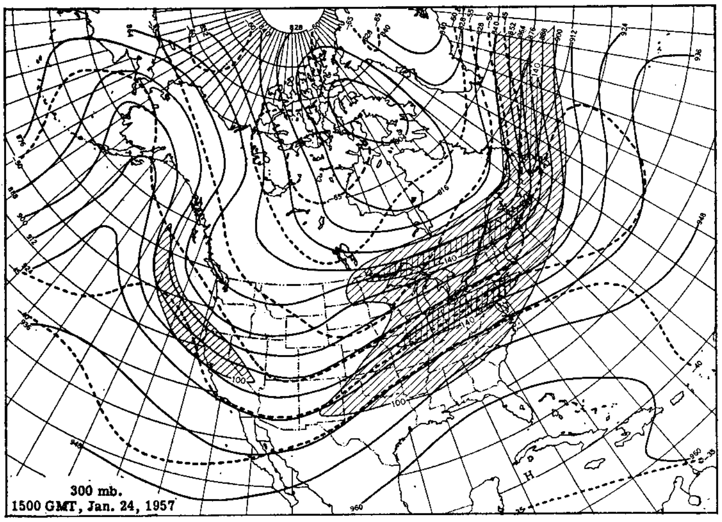

Figure 5.—300-mb. chart, 1500 GMT, January 24, 1957. Contours (solid lines) in tens of meters, isotherms (dashed lines) in degrees Celsius, and isotachs (thin solid lines) in knots. Area of winds over 100 kt. is hatched with area of maximum cross-hatched.

were about 5° C. warmer than at Salem, where on February 4, −80° C. was recorded at 33 mb.

At Peoria, Ill. (fig. 2) by January 21, the stratopause dropped into the 20- to 40-mb. layer where it remained until February 1. In the thermoincline during this period, higher temperatures were reported at Peoria than at Denver; for example, a 10-mb. temperature of −30° C. at Peoria on January 22. Figures 1 and 2 bring out the striking fact that the warming over this part of the continent was very high in the stratosphere and *was not detectable from a casual examination of temperatures at the 25-mb. surface or below.*

Figure 3, combining data for Albany, N. Y., Portland, Maine, New York City, and Nantucket, Mass., shows that the stratopause dropped into the 60- to 80-mb. layer by January 23. Although the highest valued isotherm shown above the stratopause is for −40° C., the degree of warming was not necessarily less than that at the western stations where soundings penetrated to much greater heights and therefore into warmer layers.

In searching for similarities and trends in figures 1–3, we find that a deep column of warm air associated with a mass of cold tropospheric air moving along below it moved through the stratosphere from Denver on the 16th to Albany on the 18th. Just above the 200-mb. surface a cold high tropopause appeared at Denver on the 20th, at Peoria on the 21st, and at Albany on the 22d and was surmounted by a column of cold air to a height of about 25 km. (25 mb.) at Denver, 22 km. (40 mb.) at Peoria, and 18 km. (60 mb.) at Albany. The well-marked tropopause and cold air surmounting it appeared simultaneously with a powerful thrust of warm southerly winds in the lower troposphere. That portion of the cold stratospheric air above 100 mb. was an extension of the Arctic stratospheric airmass which after its arrival remained in that region and participated only slightly in the activity at and below the tropopause. This isolation from the daily changes at lower levels was also a characteristic of the warm region above the stratopause. Once the stratopause appeared over a station, it remained at about the same level but sloped downward toward the east, so that at Albany only a thin wedge of the Arctic stratospheric airmass remained.

The vertical motion during this period can be roughly estimated for the path from Denver to Albany. The winds in the upper stratosphere were blowing at speeds from 50 to 100 kt. from Denver through Peoria to Albany. The temperature below the stratopause was fairly uniform vertically and horizontally, indicating little vertical motion in mid-stratosphere. Above the stratopause, the 25-mb. temperature was about 20° C. higher at Albany than at Denver, 1400 nautical miles upstream. With an inversion of 5° C./km., an average sinking rate of 1 to 3 cm./sec. is required along this path during the period of more than a week that the condition persisted. Higher rates occurred over shorter periods and distances (Craig and Hering [2]).

The most rapid change in the thermal pattern of the stratosphere began on January 24 in the vicinity of Goose Bay, and Stephenville, Newfoundland. Because of errors introduced by irregularities in the calculated rate of ascent of Goose Bay radiosonde observations during that period, the events are illustrated here by means of a time section for Stephenville, Newfoundland (fig. 4). This time section has been extended down to the 1000-mb. surface to include a generalized frontal pattern. The fronts did not penetrate downward to 1000 mb. in the cases of wave cyclones passing south of Newfoundland. Because time increases to the right, frontal patterns appear to be reversed, and ridges and troughs as shown by wind shifts appear to be interchanged.

Stratospheric warming occurred prior to January 20 and was followed by the appearance of a high cold tropopause with a brief incursion of Arctic stratospheric air in mid-stratosphere. Beginning on January 23, the low temperatures of mid-stratosphere modified to the more usual temperatures for this region. The stratopause settled into the 60- to 70-mb. layer, and strong warming continued above it until the 29th when a temperature of −22° C. was observed at 20 mb. Very strong winds occasionally exceeded 150 kt. during this period but decreased almost simultaneously with cessation of the warming trend.

At the commencement of the sudden or "explosive" warming following January 24, the tropospheric jet stream was unusually strong as shown by the 300-mb. chart (fig. 5). The pattern of this chart developed slowly and is quite similar to the monthly mean 300-mb. chart for January 1957 [24.] Over Canada there was a very cold Low of the type that intensifies with height up through the stratosphere, as described by Riehl [15.] A warm ridge extended northward from the Gulf of Alaska into the Arctic. Advection of cold air by the broad current of northerly winds between the Canadian Low

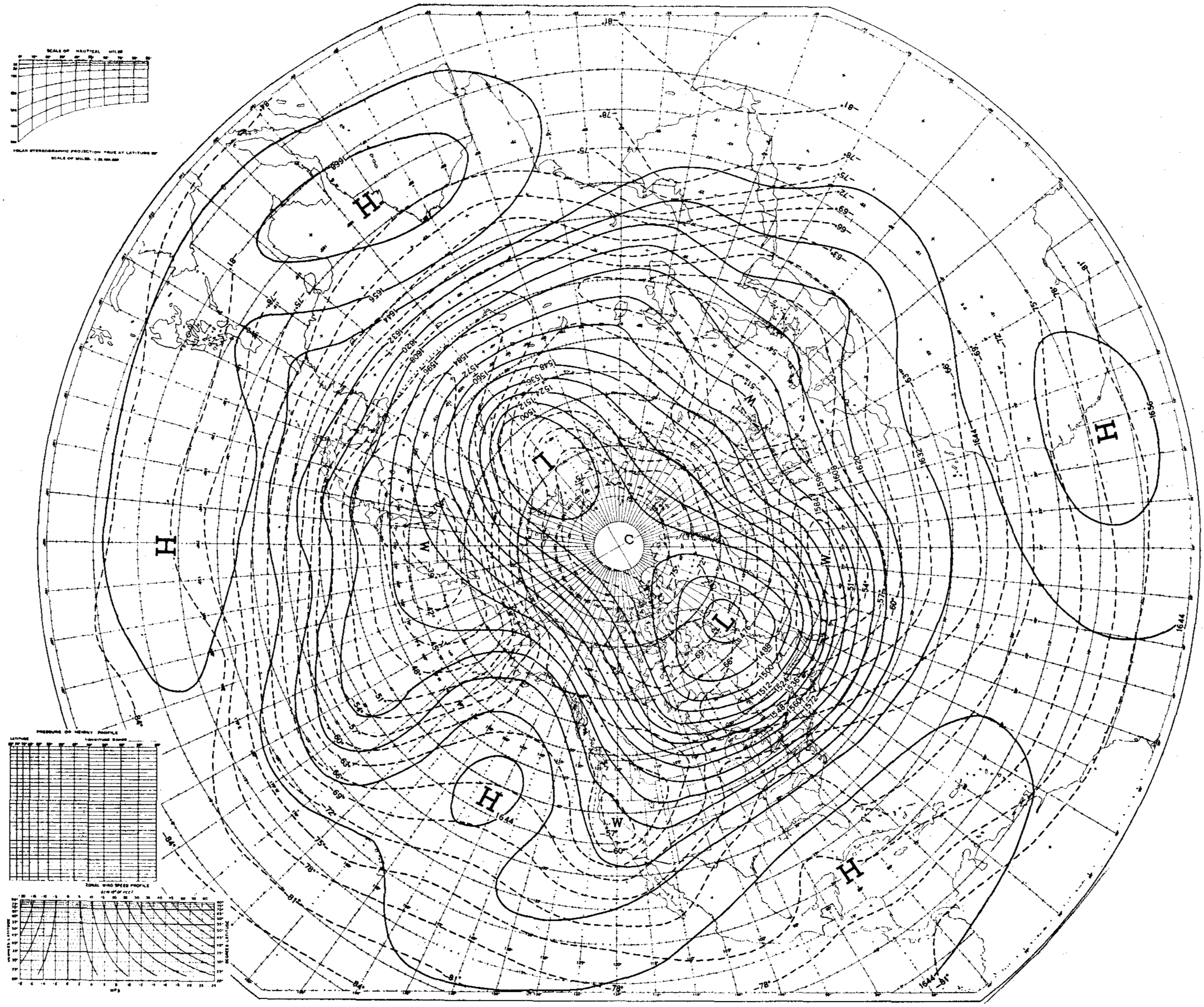

Figure 6.—100-mb. chart, January 27, 1957. Contours (in tens of meters) are shown by solid lines, isotherms (°C.) by dashed lines.

and the Alaskan ridge was an important factor in the persistence of a south-to-north temperature gradient and strong high-level westerly winds of exceptional strength across the United States. These winds accelerated as they traversed the confluence zone leading to the area of strong contour gradient between the Canadian cyclone and the ridge over Florida. The 100-kt. isotach covered the eastern United States north of 35° N., and winds exceeding 140 kt. extended in two bands from the Great Lakes to a sharp trough over the Maritime Provinces. A possible source of the kinetic energy of these winds was the potential energy released by slow settling of the inflowing cold air. In the vicinity of the sharp trough, the band of winds over 100 kt. narrowed. Since the air was moving much faster than the isotachs, it must have been decelerating and thus flowing to the right across contours.

The proposed explanation can be completed by means of reasoning used previously by Riehl and Teweles [16]. The kinetic energy released by the decelerating jet stream was reconverted into potential energy through vertical motions associated with horizontal convergence to the right of the stream and divergence to the left. In the region of divergence, the storage of potential energy can be accomplished in either of two ways, by the elevation of potentially colder air below, or the depression of potentially warmer air above, the level of zero vertical motion. Similarly in the region of convergence, the storage of potential energy may come about through the forcing of potentially warmer air downward or of potentially colder air upward. Depending upon the forces exerted in other layers, different combinations of these alternatives may predominate. Moreover, the area of maximum convergence or divergence, and thus also the area of maximum warming or cooling, may have a pro-

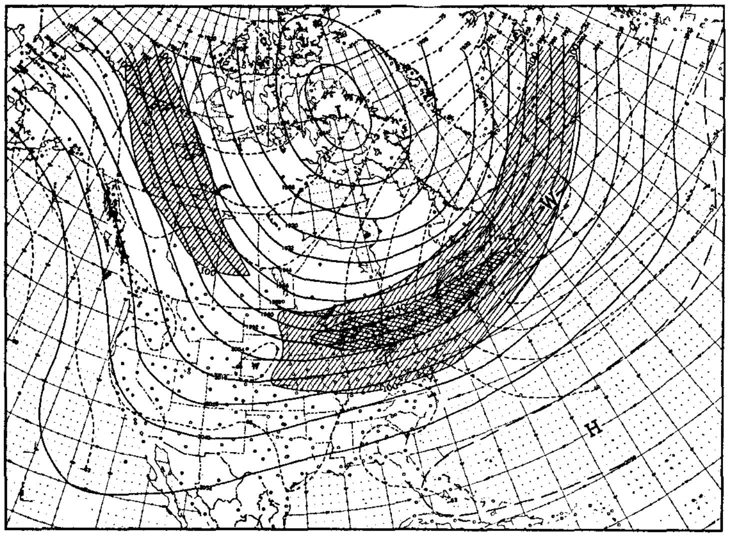

Figure 7.—50-mb. chart, January 27, 1957. Contours (in tens of meters) are shown by solid lines, isotherms (°C.) by dashed lines. Areas with winds above 100 kt. are hatched, with areas of maximum cross-hatched.

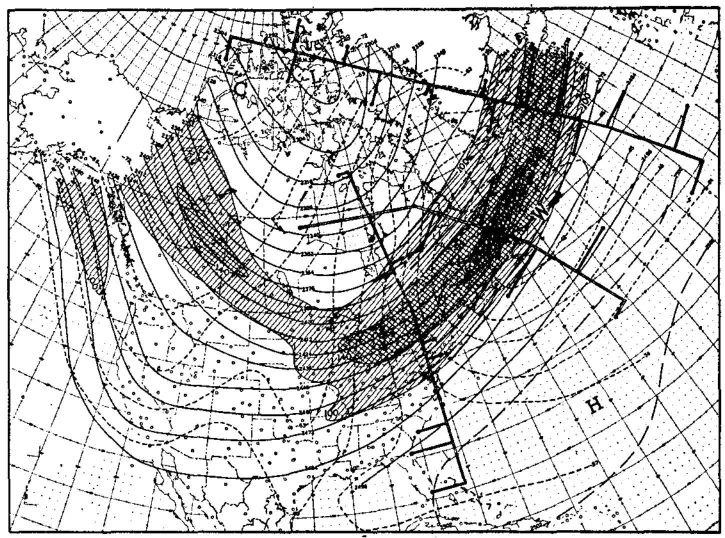

Figure 8.—25-mb. chart, January 27, 1957. Contours (tens of meters) are shown by solid lines. Isotherms (dashed lines) are means for the 50- to 25-mb. layer. Lines of intersection of the 25-mb. surface and the cross-sections in figures 10, 11, and 12 are shown, with dots locating nearby radiosonde stations. Hatching shows areas of wind above 100 kt.

nounced slope in the vertical, so that the warmest point at one level may be displaced a thousand miles or more from the warmest point at a higher level.

4. THE CIRCULATION DURING THE INTENSIFICATION STAGE

As the 25-mb. warm center moved northward from its position south of Newfoundland on January 25 (see fig. 25), its temperature increased rapidly. Various charts have been prepared to illustrate conditions at the time maximum warming was initiated. Since the warm center moved quite slowly from January 26 to 28, data for more than one observation were combined in the various charts centered at about 1500 GMT, January 27 to compensate for the scarcity of data at any one observation.

A hemispheric 100-mb. chart for January 27 (fig. 6) is presented to place events over North America in full perspective with the rest of the hemisphere. A hemispheric 50-mb. or 25-mb. chart would be preferable since the warming was more intense and clearly defined at higher altitudes, but insufficient data were available for these surfaces over Eurasia. At 100 mb., a warm center of about −50° C. was situated over the mid-Atlantic where it regenerated after moving eastward into that region on the 21st. The dual nature of this warm center is shown in the cross section from the Azores across Greenland to the Canadian Arctic (fig. 12). In the troposphere north of the Azores was the typical pattern of the polar front with its jet stream and tropopause structure. It can be seen that the warm center at 100 mb. (fig. 6) was associated in the vertical with the warm belt in the lower stratosphere north of the tropospheric jet stream, and at the same time with the strong warming of the upper stratosphere then taking place in an area centered to the northwest of this region.

From figure 6 it was discovered that a concurrent warming was taking place over stations in eastern Siberia, precisely on the opposite side of the pole. Almost all features of the phenomenon over Siberia were very similar to those over North America. Near 65° N., a deep cyclone tapped the Arctic cold pool to carry cold air into mid-latitudes. There was a confluence of contours into the region between the Low and a High centered over Burma. Between Japan and Kamchatka where the contours diverged the maximum reported temperature was 12° C. higher than in the Atlantic. Although temperature corrections may not have been applied to Russian observations at this time, the radiational temperature corrections at 100 mb. would probably be 3° C. or less. In any case, the reported temperature of −43° C. at Attu in the Aleutians substantiated the greater warmth at 100 mb. over eastern Asia.

Besides the two marked warm centers already discussed, there were at least two others at 100 mb. on January 27, one over the western United States and the other over central Europe. All four of these warm centers were situated over complex tropospheric low pressure systems, and lobes in the isothermal patterns surrounding these warm centers covered individual polar front cyclones. The effect of the deep warm layers of the lower stratosphere was to erase all trace of sea level low pressure systems from the same areas in the upper stratosphere. The two large cold, low centers that show up over Canada and Siberia on the 100-mb. chart were both located directly over the two coldest sea level areas of the Northern Hemisphere, a fact suggesting that the distribution of long-wave radiation from the surface of the earth may be a factor in determining the position of stratospheric Lows.

Conditions at the 50-, 25-, and 15-mb. surfaces in the

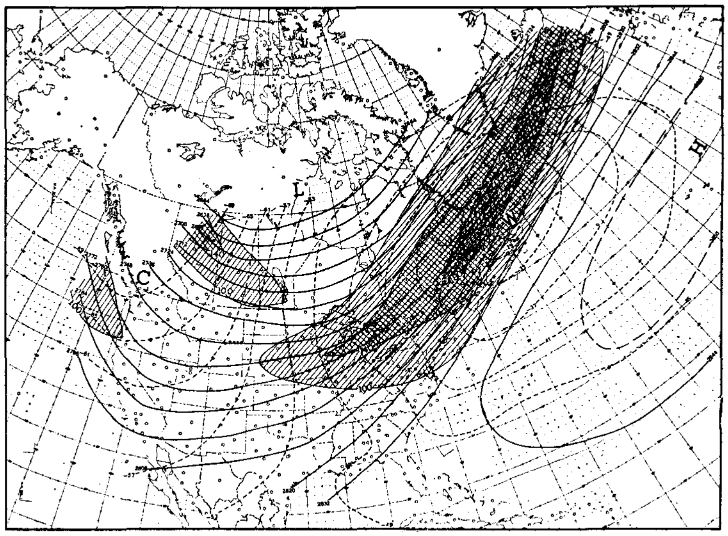

Figure 9.—15-mb. chart, January 27, 1957. Contours (tens of meters) are shown by solid lines. Isotherms (dashed lines) are means for the 25- to 15-mb. layer. Hatching shows areas of wind above 100 kt.

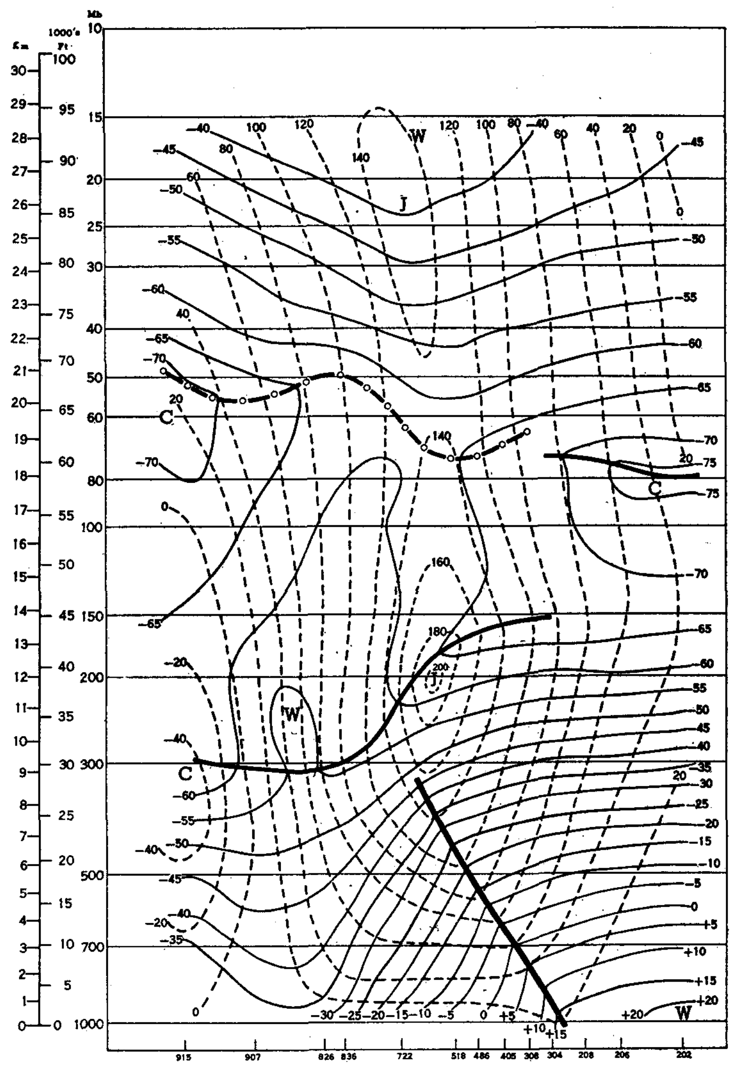

Figure 10.—Cross section for January 27, 1957, from Miami, Fla. (202) to Coral Harbour, N. W. T. (915) as delineated in figure 8. Analysis includes front (very heavy line), tropopause (heavy line), stratopause (heavy line with circles), isotherms (thin lines), and isotachs (dashed lines). Core of westerly jet stream shown by letter J.

upper stratosphere over the North American sector on January 27 are shown in figures 7, 8, and 9.[2] The mean isotherms for the 15- to 25-mb. layer (fig. 9) show very cold air over the southwestern United States where at 100 mb. there was a warm center. On the other hand, the 100-mb. warm center over the Atlantic was surmounted by much warmer air in the 15- to 25-mb. layer. The resulting contour-isotherm relationship in the upper stratosphere agreed with the observation by Wexler and Moreland [30] that troughs of the upper stratosphere are cold in their southwest quadrants and warm in their southeast quadrants. This vertical variation of the isotherm pattern above 100 mb. also substantiates the conclusion that strong effects of vertical motion in migratory tropospheric systems extend up to the 100-mb. surface, but above that surface the motions of stratospheric systems dominate.

The westward displacement and temperature increase of the Atlantic warm center upward from 100 mb. is clearly seen in figures 7–9. Of interest also is the relation between strongest winds at the 50-mb. and 25-mb. surfaces, and the warm center in the layers above and below. In the drawing of isotachs at these surfaces, the anticyclonic shear south of the jet axis was deliberately held short of the criterion for dynamic instability even though the shear based on the construction of contours by use of mean temperature charts tended to exceed the criterion. The winds, approaching the warm center at the 50-mb. surface and in the layer below, were decelerating. Also if we assume adiabatic temperature changes, these winds descended as they moved into this region of depressed isentropic surfaces. Thus we have a current in which decrease of kinetic energy took place as the current passed through a region where buoyancy not only increased downstream but also was increasing with time as the isentropic surfaces become farther depressed. The kinetic energy apparently was being converted into potential energy as represented by the increased tilt of the isentropic surfaces surrounding the warm center. The isotach-isotherm pattern changed with height so that at the 25-mb. and 15-mb. surfaces the center of strongest winds was very close to the warm center, a relationship to be expected if the warm center and thus the strong temperature gradient northward from the center were produced by forced sinking into a lower layer of divergence. The vertical slope of the warm center into the upstream direction and toward the low pressure side of the current also agreed with the concept of a current flowing to the right across contours to produce convergence with sinking in lower layers to the right of the stream and divergence with sinking in higher layers to the left of the stream.

[2] Even though almost endless consistency checks were made to overcome deficiencies in the data for these and other charts for the very high altitudes, inaccuracies probably remain, and reliance should not be placed in minor details and small differences.

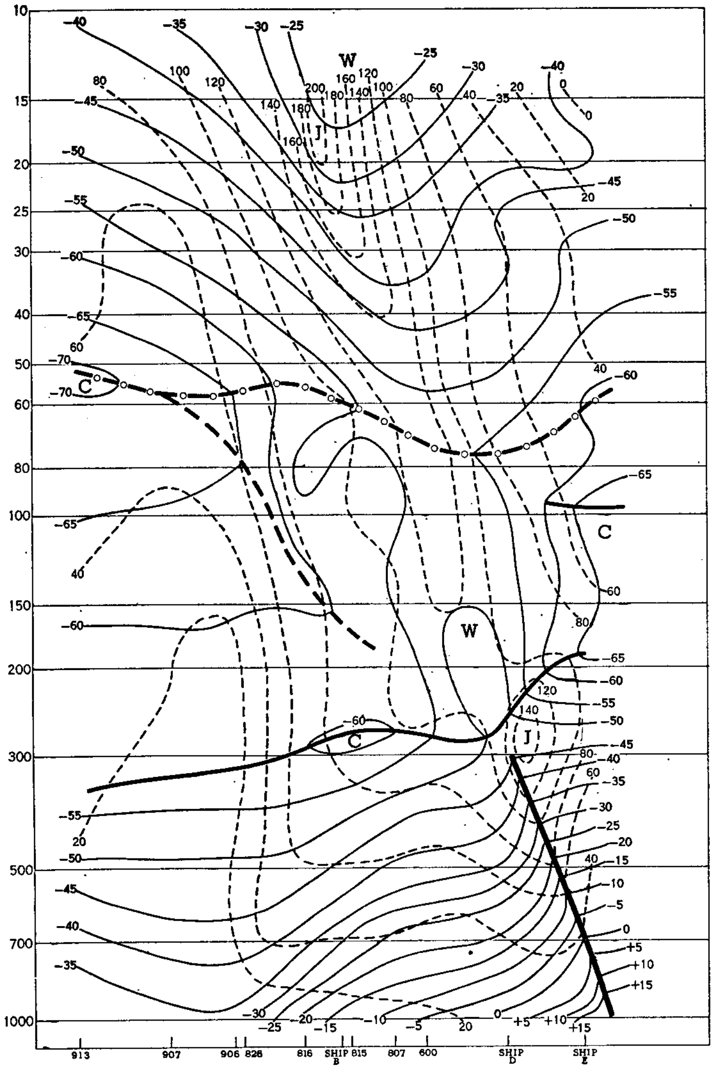

Figure 11.—Cross section for January 27, 1957, from Atlantic Ship E to Churchill, Manitoba (913) as delineated in figure 8. Analysis as in figure 10 but with lapse rate discontinuity shown by heavy dashed line.

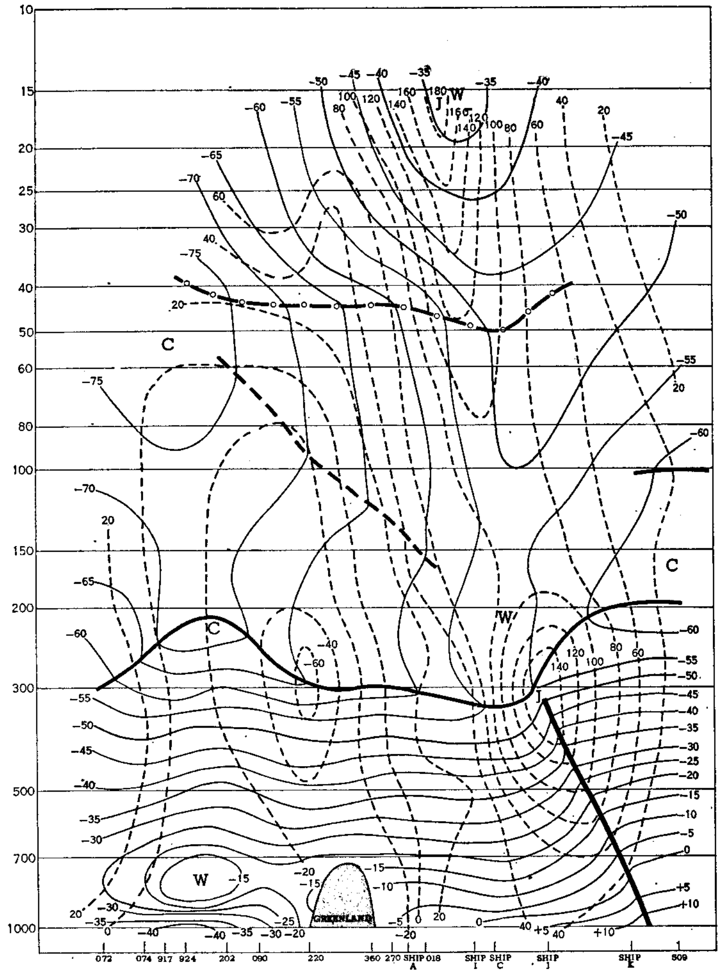

Figure 12.—Cross section for January 27, 1957, from Lajes, Azores (509) to Mould Bay, N. W. T. (072) as delineated in figure 8. Analysis as in figure 11.

To illustrate the interrelationship of wind flow and temperature field in greater detail, three cross sections (figs. 10, 11, and 12) were constructed nearly perpendicular to the wind flow at the higher levels and along lines shown in figure 8. Vertical slopes of all features are greatly exaggerated by the large distortion of the vertical scale, about 200 to 1. Godson and Lee [6] used similar cross sections in their description of this and earlier cases and also in their description (Lee and Godson [10]) of the Arctic stratospheric jet stream over Canada during the winter of 1955–1956. The cross sections shown here include the more prominent tropopauses, to aid in distinguishing tropospheric airmasses from those of the stratosphere.

At least four stratospheric airmasses can be identified although their boundaries, drawn at the cold side of the transition zone, were sometimes indistinct. Above the tropical tropopause, usually found in or near the 80- to 100-mb. layer, there was the *tropical stratospheric airmass with moderate temperature inversion.* Above and north of the tropospheric jet stream was the *warm isothermal stratospheric airmass of the high mid-latitudes.* To the south of this airmass between 100 and 200 mb., temperature decreased toward the high cold tropical troposphere, and to the north temperature decreased toward the *Arctic stratospheric airmass* that forms in the darkness of the polar night. There is generally a slight lapse rate in the latter airmass, but soundings taken on its periphery sometimes show an upper boundary above which there is an isothermal lapse rate or inversion. In cases of anomalous warming, sinking motion in the upper stratosphere produces a *superior stratospheric airmass,* characterized by an inversion of 3° C. per km. or more and separated from the lower layers by the lapse rate discontinuity we refer to as the stratopause.

The geometrical relationship of the various features across a wide span of longitude is illustrated in figures 10, 11, and 12. These relationships may not be typical of all anomalous warmings since, as Godson and Lee [6] have demonstrated, there were remarkable dissimilarities among anomalous warmings that have been studied. Near 80° W.

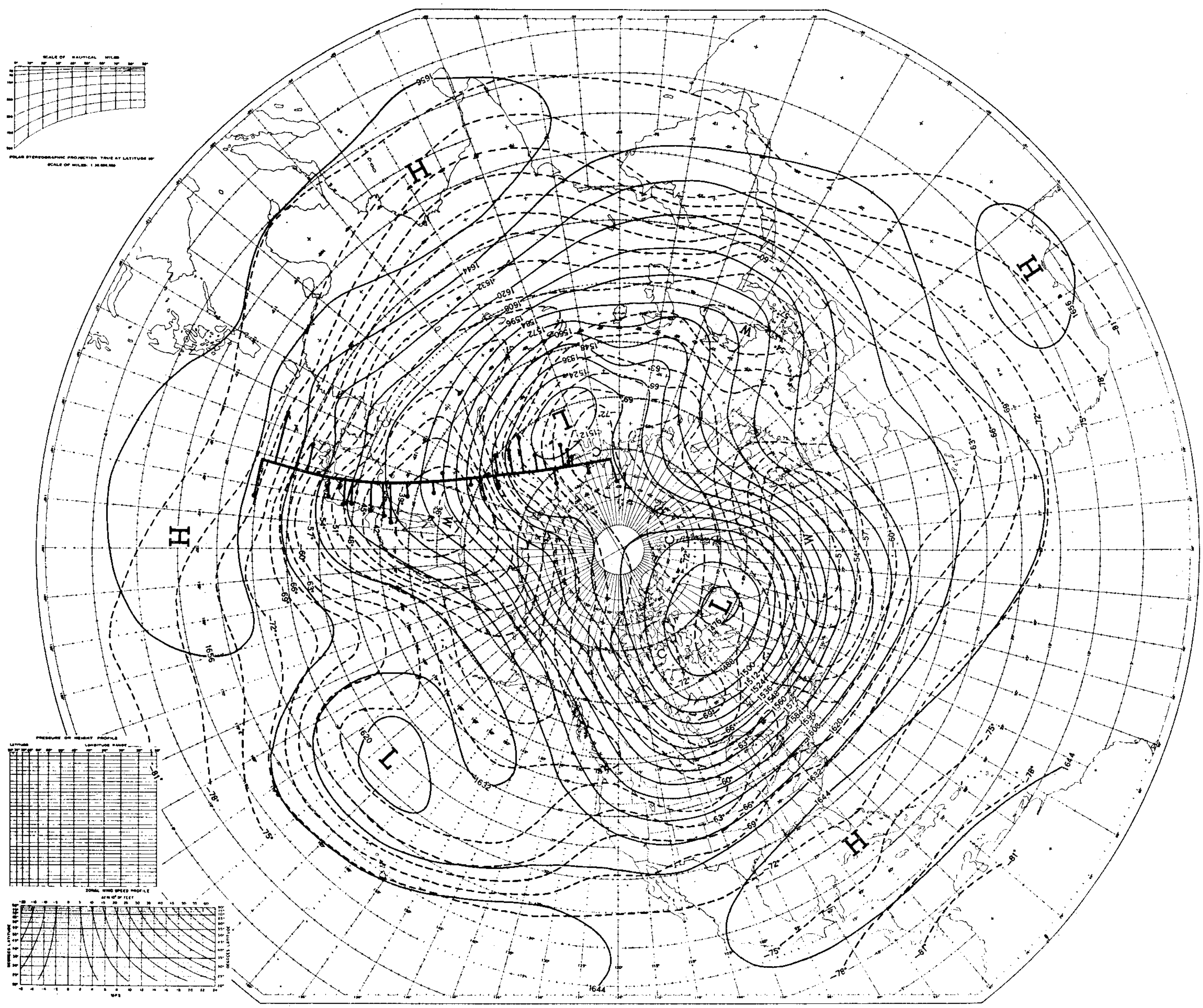

Figure 13.—100-mb. chart, February 1, 1957. Contours (tens of meters) are shown by solid lines, isotherms (°C.) by dashed lines. Line of cross section in figure 21 is shown by heavy line with dots locating nearby radiosonde stations.

(fig. 10) the jet stream of the polar front was exceptionally strong with speeds exceeding 200 kt. near 200 mb. The usual large decrease of speed above the jet core was greatly reduced here since the mid-latitude warm belt was pinched off by intrusion of the Arctic stratospheric airmass. Thus high wind speeds extended high into the stratosphere. The temperature gradient northward from the region of warmest stratospheric air and the associated vertical increase of westerly wind component were thus superimposed upon the belt of strong westerly winds and accounted for the upper stratospheric jet stream that appears near the top of the cross sections. In the cross section taken nearly through the center of warmest stratospheric air (fig. 11), the polar front jet stream was much weaker than farther west (fig. 10), but the stratospheric jet stream attained maximum strength in agreement with the strong temperature gradient northward from the center of maximum temperature.

The area of highest 100-mb. temperature in the easternmost cross section (fig. 12) was not surmounted by a strong inversion in the upper stratosphere, and the past movement of this 100-mb. warm center had been closely associated with the eastward-penetrating polar front jet stream. In this cross section, maximum temperatures at surfaces above 50 mb. were not as high as they were farther west (fig. 11) and all warming would be adequately explained by advection. Figure 12 extends far into the Arctic to show the structure of the stratosphere in the region of minimum temperature where because of the very low temperatures, radiosondes did not penetrate to the

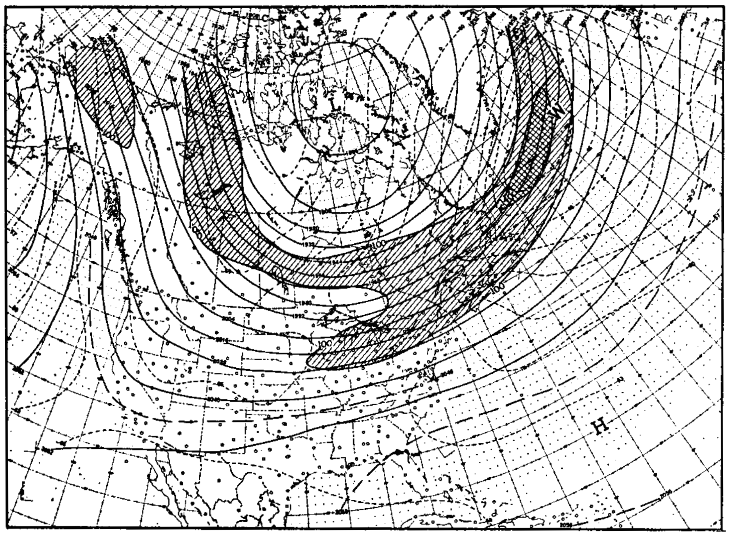

Figure 14.—50-mb. chart, February 1, 1957. Contours (tens of meters) are shown by heavy solid lines with intermediate lines dashed; isotherms (°C.) are thin dashed lines. Hatching shows areas of wind above 100 kt.

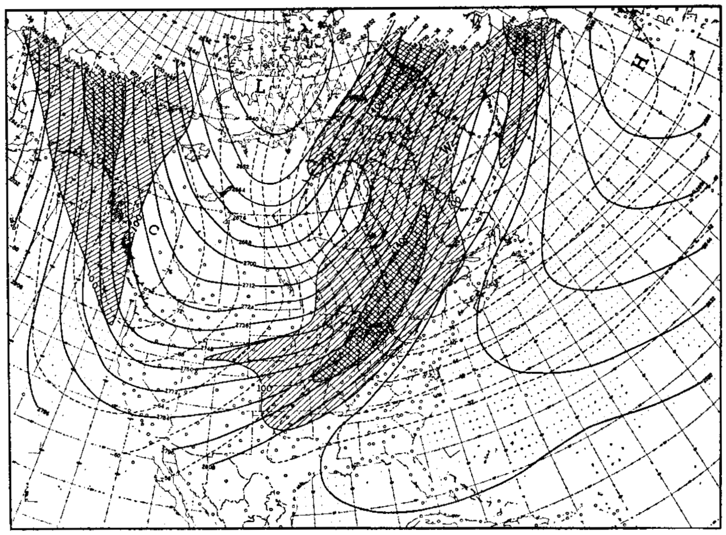

Figure 16.—15-mb. chart, February 1, 1957. Solid lines are contours, dashed lines mean isotherms for 25- to 15-mb. layer. Hatching shows areas of winds over 100 kt.

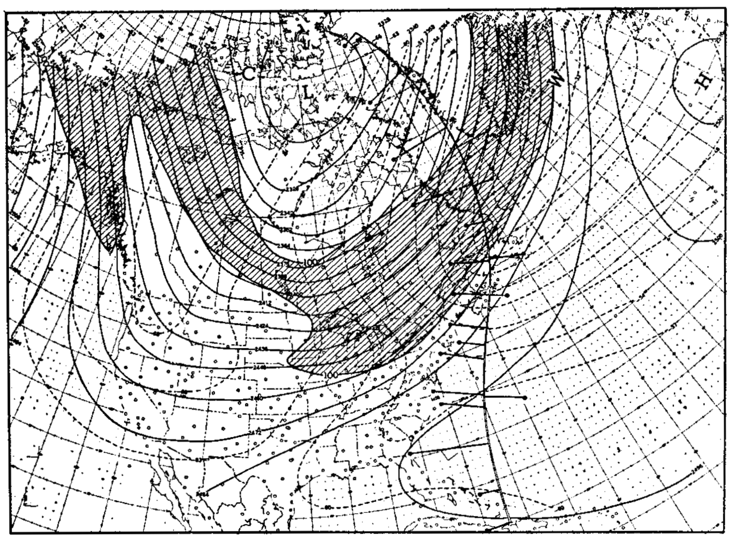

Figure 15.—25-mb. chart, February 1, 1957. Contours (tens of meters) are shown by solid lines. Isotherms (dashed lines) are means for the 50- to 25-mb. layer. Line of cross section in figure 19 is shown with dots locating nearby radiosonde stations.

greatest heights. The elevation of the Arctic tropopause to near the 200-mb. surface is an interesting feature of this cross section and confirms the observation, made by Riehl [15], of an exceptionally high Arctic tropopause in the vicinity of a cold stratospheric Low. A similar feature in the Antarctic was described by Schumacher [18].

5. THE MIGRATORY STAGE

Five days later, on February 1, the rate of intensification of the upper stratospheric warm center had diminished, but its northward movement had accelerated. Lower down in mid-stratosphere at 100 mb. (fig. 13), the Atlantic warm center had moved slowly toward the northeast between Greenland and the British Isles. Near the North Pole, continued increase of meridional circulation had permitted modification of the temperature with disappearance of the −75° C. isotherm and splitting of the −72° C. isotherm. The 100-mb. warm center over the east coast of Siberia had intensified and moved northwestward perpendicularly to the current. During the same 5-day period, marked retrogression at sea level and 500 mb. took place over the entire northern Pacific.

The upper stratospheric constant pressure charts for the North American sector (figs. 14, 15, and 16) clearly show the intensification and northwestward displacement of the warm center with height. The stratospheric jet stream was in nearly the same relation to the warm center as 5 days earlier except at the highest layer between 15 and 25 mb. where the warm air had progressively overspread the jet stream of the layers below to cause its decay with height as shown by the relatively disorganized 15-mb. isotach pattern. Over the Alaskan area on the western side of the trough, wind speeds increased with height up to the 15-mb. surface. However, soundings to above the 10-mb. surface at Salem, Oreg., and Spokane, Wash., showed a marked inversion above 15 mb. Radiosonde observations at western Alaskan stations to the right of the current did not extend above 15 mb. but showed only continued cooling with height to the top of the soundings. If there was no similar inversion to the right of the current, then the temperature gradient and vertical wind shear had become reversed above 15 mb., and the upper stratospheric jet stream in the area had reached its maximum at the 15-mb. surface.

The synoptic pattern in the upper stratosphere during this 5-day period changed very little over the North American continent, and even over the Atlantic there was little change up to the 50-mb. surface, but the rapid temperature rise above that surface in mid-latitudes was

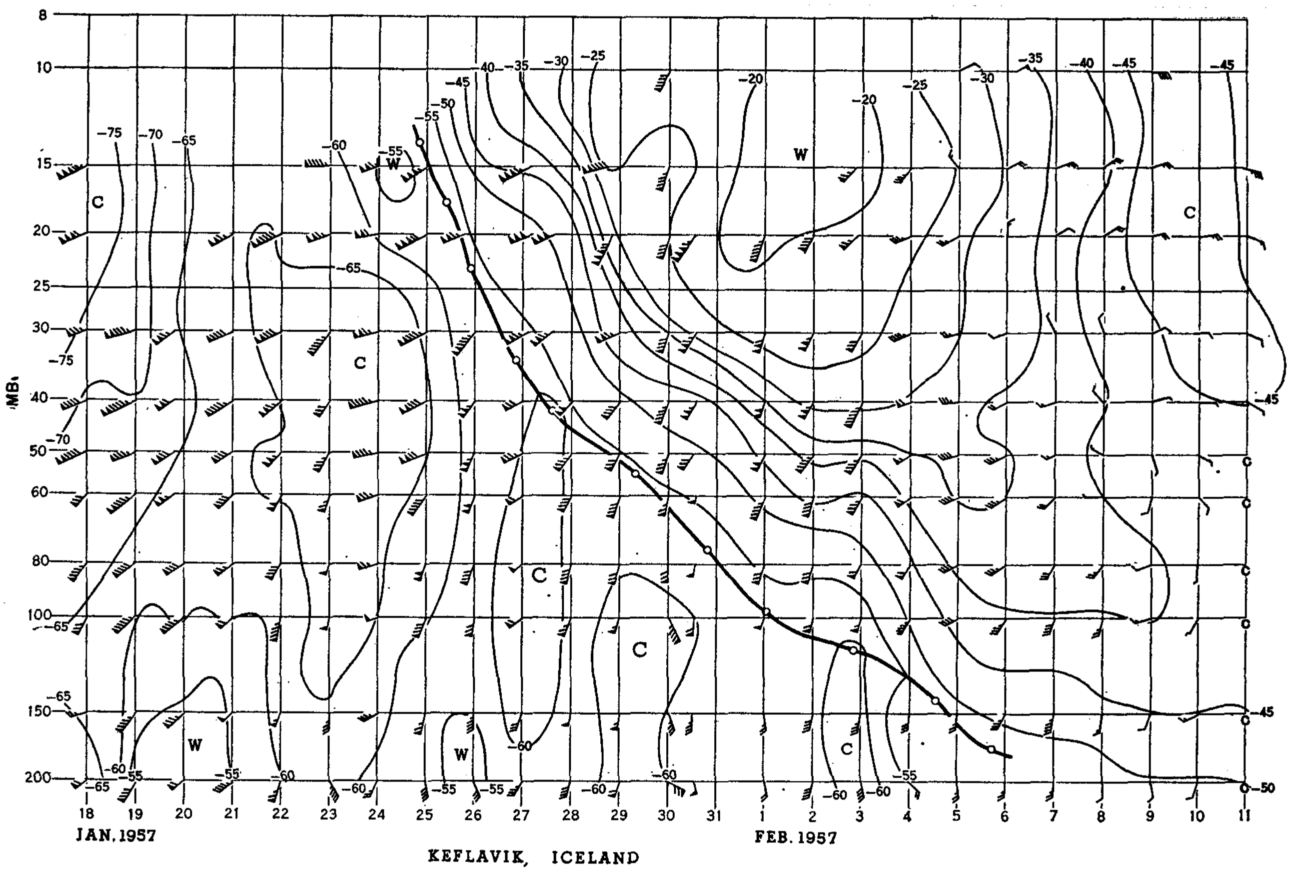

Figure 17.—Time section, January 18 to February 11, 1957, for Keflavik, Iceland. Isotherms are in °C., winds oriented from top of diagram for north. Heavy line shows stratopause. Vertical grid lines are for 1500 GMT of dates shown.

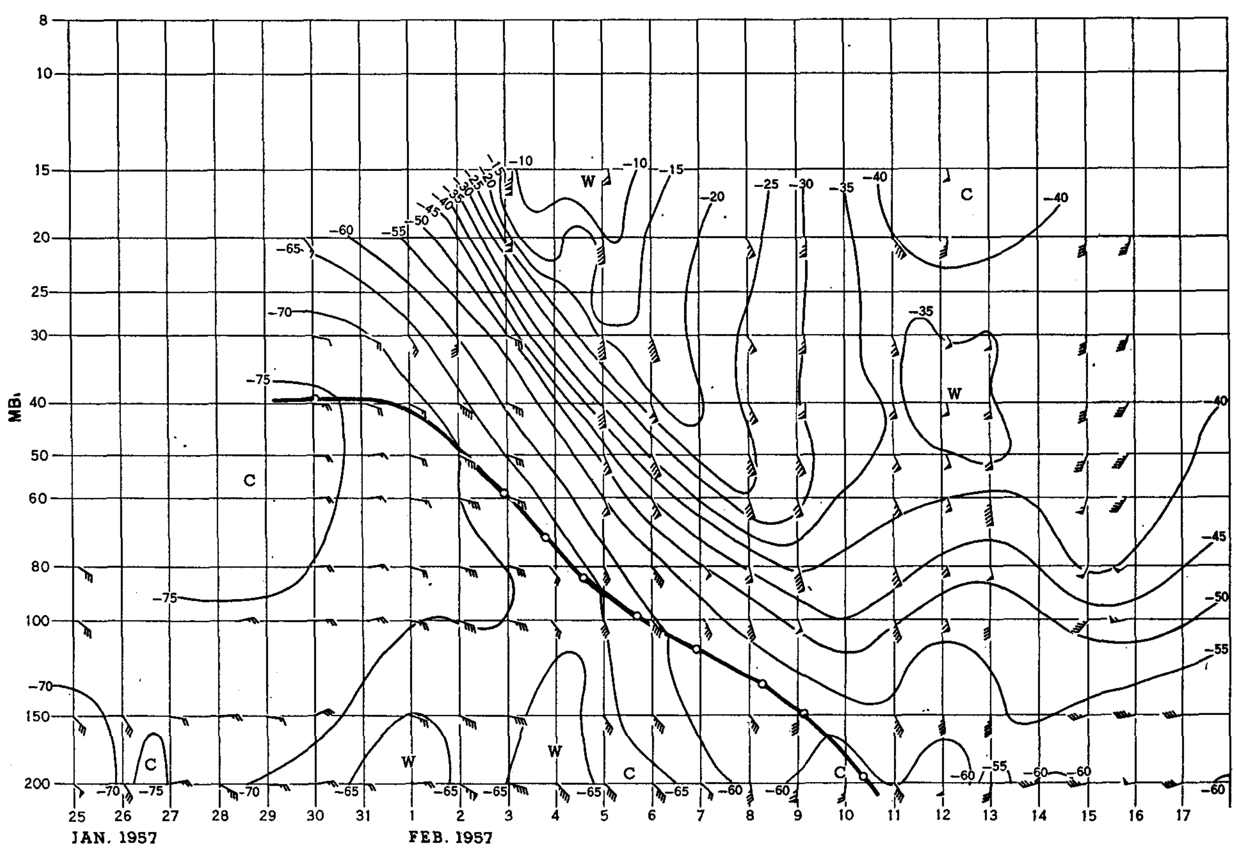

Figure 18.—Time section, January 25 to February 17, 1957, for Thule, Greenland. Isotherms are in °C., winds oriented from top of diagram for north. Heavy line shows stratopause. Vertical grid lines are for 1500 GMT of dates shown.

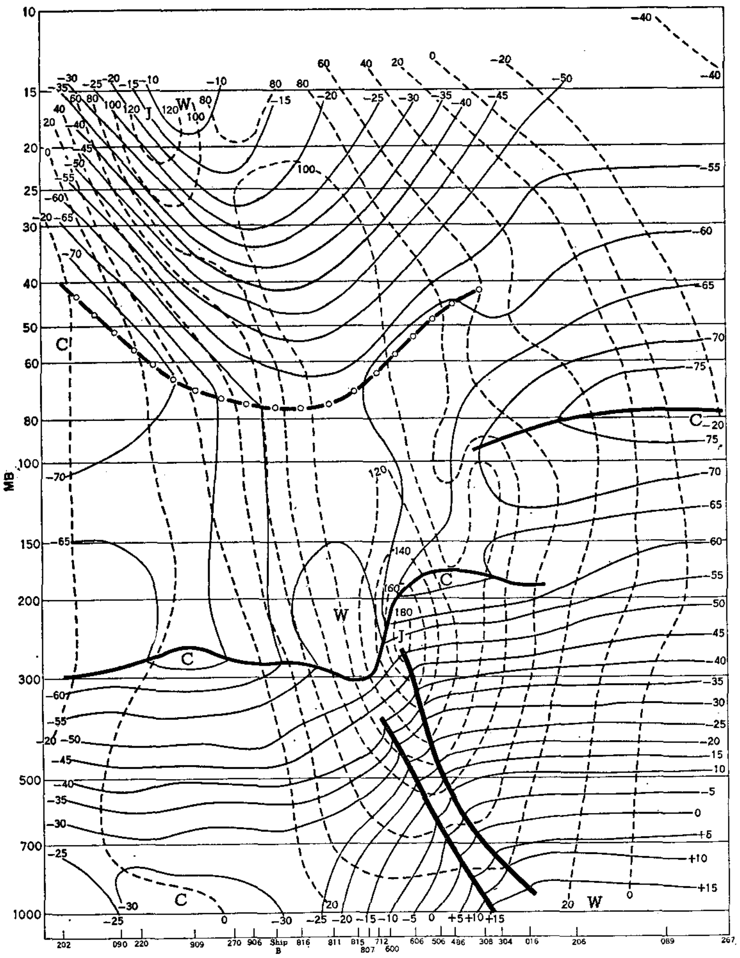

Figure 19.—Cross section for February 1, 1957, from Guantanamo, Cuba (267), to Thule, Greenland (202), as *delineated in figure* 15. Fronts are shown by very heavy lines, stratopause by heavy line with open circles, tropopause by heavy line without circles. Isotherms are thin solid lines, isotachs, dashed. Core of westerly jet stream is shown by J.

associated with equally rapid anticyclogenesis, particularly south of Iceland, as may be seen by comparing figures 9 and 16. The time section for Keflavik, Iceland (fig. 17) at 64° N. is markedly different from those for stations farther west (figs. 1–4). At this high latitude during the first half of January, temperatures above the 50-mb. surface were frequently lower than −75° C., falling below −80° C. in at least one instance. As wind speeds increased to more than 150 kt., a slow but irregular warming of the upper stratosphere began on January 18. After January 26 the rate of warming was particularly rapid up to February 1 when −15° was reported at 20 mb. An inspection of wind reports plotted in figure 17 shows that the wind speed decreased abruptly at each level as the warm axis at that level passed northward across the station to reverse the temperature gradient in the layer below. Warming also took place very rapidly at Narsarssuak and Thule, Greenland (fig. 18) and at Arctic stations in Canada. The higher levels began to warm first and at some stations reached maximum temperature before warming at lower levels was even initiated. The total increase of temperature seems to have been greatest in the 10- to 15-mb. layer, although this cannot be fully verified since temperature observations above 15 mb. were rare prior to the warming due to the extreme cold and during the warming due to increased wind speeds.

A cross section for February 1 (fig. 19), taken along the line drawn from Cuba to northern Greenland in figure 15, cuts through the center of warmest upper stratospheric temperatures. The warmest 15-mb. temperature was fully 15° C. warmer than that in the analogous section for January 27 (fig. 11). During the last part of the 5-day period the center of warm air accelerated toward the north, thereby increasing its distance from the tropospheric jet stream and other mid-latitude features and moving over colder air in the lower stratosphere. The temperature gradient north of the warm center became situated over a region of very light winds so that the associated jet stream in the upper stratosphere did not reach the high speeds found on previous days. Warming of the lower stratosphere amounted to only 2° or so in a day and could have been a radiational effect of the rapidly warming upper layers.

The simultaneous warming that occurred in the southeastern portion of the trough located over Siberia is detailed in a time section for Yakutsk at 74° N., 130° E. (fig. 20) and in a cross section from Iwo Jima to Novaya Zemlya (fig. 21), taken along the line shown in figure 13. At 80 mb. above Yakutsk, a total temperature increase of 30° C. occurred between January 24 and February 5. Strong winds were observed up to 100 mb. during the warming. The horizontal temperature gradient indicated higher speeds at higher levels. A stratopause was hardly detectable at Yakutsk where the lower stratosphere simultaneously participated in the warming almost as strongly as the upper stratosphere. The cross section for 0200 GMT, February 1 (fig. 21) agrees with this feature. The temperature of the lower stratosphere just north of the tropospheric jet stream was 10° C. or more warmer than the analogous temperature in the Atlantic area. The contrast between abnormally warm stratosphere in middle latitudes and extreme cold in both the Arctic and the Tropics, as well as the different structures of the cold regions, is clearly revealed in figure 21. In the Tropics, the cold air appeared only as a thin wedge at the tropopause (see also [9] and [13]) while in the Arctic the cold air extended as high as the radiosondes penetrated.

It is clear from the fairly light winds and variable shear in the region of the warmest air in figure 21 that the exceptional warming cannot be explained by action of the wind just in the neighborhood of the cross section. A clue to the origin of warming is given in figure 13 which shows that the air warmed steadily as it decelerated in traversing the quasi-stationary region of isobaric difluence that extended eastward from the constriction of the isobars between the High over the Bay of Bengal and the Low in northern Siberia.

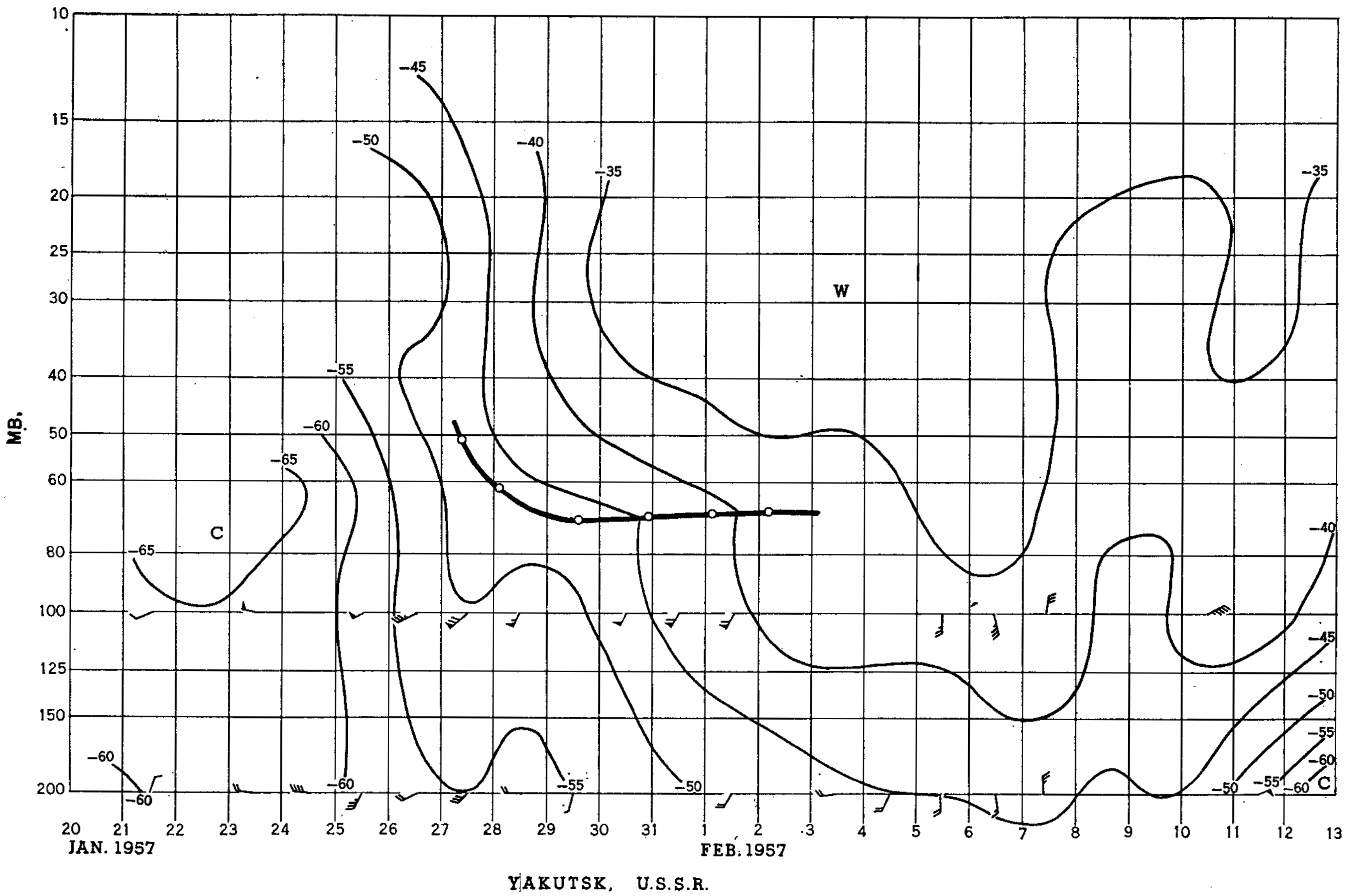

Figure 20.—Time section, January 21 to February 12, 1957, for Yakutsk, U.S.S.R. Isotherms are in °C.; winds oriented from top of diagram for north. Heavy line is stratopause. Vertical grid lines are 1500 GMT on dates shown.

6. THE MATURE STAGE

A glimpse of the warm area over North America during the mature stage is afforded by the 25-mb. chart for February 6 (fig. 22). Around the low center then moving southwestward across Canada, there was a nearly uniform vortex of winds of about 100 kt., in contrast to the jet stream with marked maximum speed that appeared on January 27 (fig. 8) and February 1 (fig. 15). The cold air off the west coast of the continent was completely pinched off from its Arctic source region, but the minimum temperature at 25 mb. remained almost unchanged in spite of the large increase of temperature in the other quadrants of the Low. Continued anticyclogenesis over the Atlantic culminated in a closed High, moving toward the northwest. The magnitudes of the height rises at various levels as the High passed over Iceland can be compared by means of a graph of the heights of constant pressure surfaces at Keflavik (fig. 23). During the first half of January there was close correspondence between height changes of all constant pressure surfaces with a slight vertical increase of amplitude to indicate slight warming accompanying the height rises. After January 23 the 15-mb. surface began to rise and with it the 30-mb. surface, but at a slower rate. The 100-mb. surface rose briefly but then settled back again until February 2 when a strong, steady rise began. The amplitude of rises was much greater at the upper levels, amounting to 2,500 m., or 8,000 ft., at 15 mb. in the two weeks following January 23 and ranging over a full 3,300 m. or 10,800 ft. at 10 mb. during January and February. The percentage pressure change is indicated by the fact that the 15-mb. surface at Keflavik on several days in early February was at a greater height than the 10-mb. surface had been on January 5.

A view of the vertical temperature distribution on February 6 through the Atlantic High and Canadian Low is afforded by a cross section (fig. 24), taken along a curved line (fig. 22) from the Azores to Oregon. The −5° C. isotherm concurs with the observation of −3.4° C. at 11 mb. at Churchill 12 hours earlier. This observation, listed in the Northern Hemisphere Data Tabulations [25] was discussed by Lowenthal [11] who also mentioned an unlisted report of −3.5° C. at 7 mb. on February 6 at Salem. This latter report would indicate that the warming effect at very high levels had spread almost completely over the coldest air in mid-stratosphere and that the −5° C. isotherm should be open at the top. The maintenance of low temperature near 50 mb. in the same atmospheric column with far warmer air below and above is remarkable and requires an explanation through dynamic process since the isotherms everywhere moved much more slowly than the winds. Over the Atlantic stratospheric warming

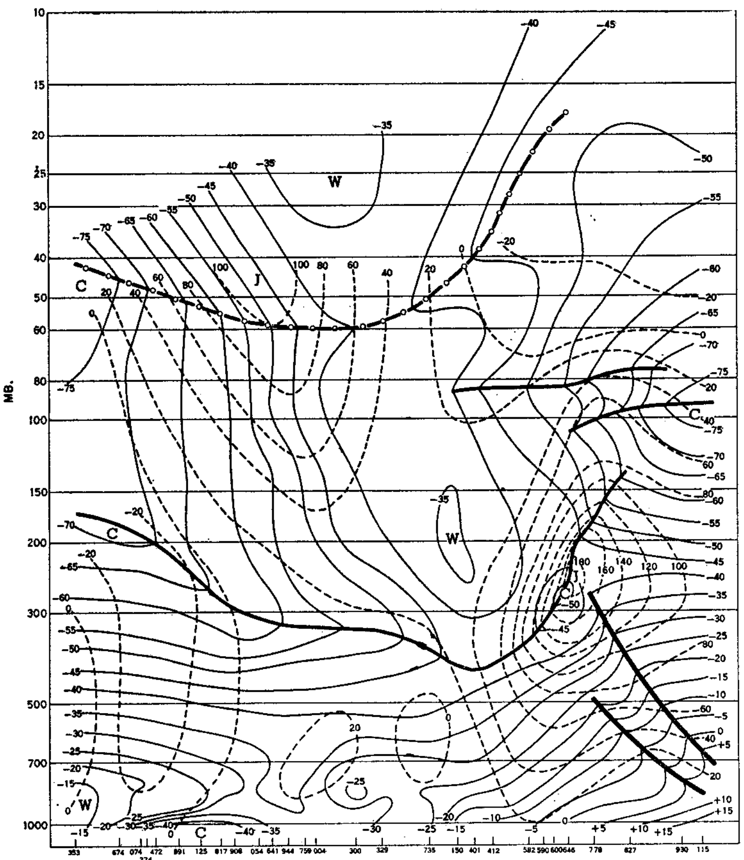

Figure 21.—Cross section for February 1, 1957, from Iwo Jima (115) to Cape Zhelaniya, Novaya Zemlya (353) as delineated in figure 13. Fronts are shown by very heavy lines, stratopause by heavy line with open circles, tropopause by heavy line without circles. Isotherms are solid lines, isotachs dashed. J shows center of westerly jet core.

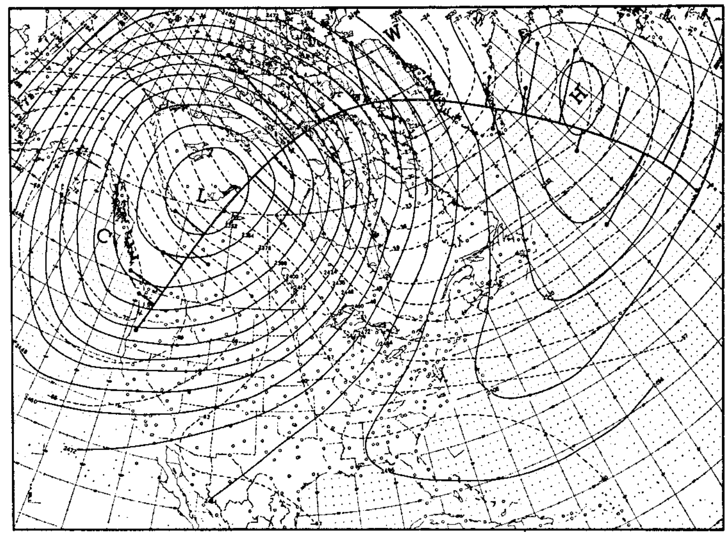

Figure 22.—25-mb. chart, February 6, 1957. Contours (tens of meters) are solid lines, isotherms (°C.) dashed. Line of cross section in figure 24 is shown with dots locating nearby radiosonde stations.

penetrated downward until the stratopause merged with the tropopause. Southward from the sloping axis of warmest air in the stratosphere, a nearly isothermal condition was achieved as temperatures returned to more seasonable values.

To illustrate other aspects of the warming phenomenon, the tracks of the warm and cold centers at 25 mb. are shown in figure 25 (adapted from Thompson [21]). On January 16, a warm center became definable on the United States-Canadian border and moved eastward for four days in a direction nearly that of the wind blowing through it but at a much slower speed. During this early stage, the 25-mb. warm center, displaying vertical linkage through an exceptionally deep layer (see also figs. 1–4), moved along directly above a very cold, intense trough in mid-troposphere. This is illustrated by the time section for Stephenville (fig. 4) where on January 19 and 20 the passage of the warm stratospheric air was accompanied by a low tropopause surmounting the very cold tropospheric air. During the diffuse stage from January 20 to January 25, the warm mass of air at 25 mb. spread amorphously over a portion of the eastern Atlantic, and the position of the center was uncertain. On January 22, the center came to a halt and began to retrogress against the wind flow. Its central temperature rose slowly at first but more rapidly, from −41° C. to −32° C., during the first 24 hours of the intensification stage beginning on January 25, and to −18° C. by January 29. Movement of the center was nearly perpendicularly to the left of the wind during this stage and until at least February 10.

On the western side of the continent, the cold center, with temperatures as much as 68° C. lower than in the warm center, could be definitely located only on the four days from February 1 to 4. It also moved nearly perpendicularly to the wind but to the right. Examination of the 25-mb. charts at 24-hour intervals reveals an interesting parallelism of trajectories of cold center, cyclone center, warm center, and anticyclone center, one following almost in the tracks of the other with warm center and anticyclone developing as cold center faded and cyclone center filled.

7. INTERPRETATION AND CONCLUSIONS

At this time it is logical for the reader to ask how the 1957 warming epoch differed from the usual behavior of the stratospheric circulation and whether the circulation and its variations can be explained in the same way as their tropospheric counterparts. It would be unwise to draw general conclusions solely from the circumstances attending a single exceptional case. Yet a case having features of heroic proportions is most likely to suggest profitable avenues of investigation, and the search in this instance is facilitated by several types of source material already in the literature. From published mean values [9, 14, 22] anomalies can be approximated. Several case studies of the circulation patterns accompanying the annual departure of the Arctic stratospheric airmass from the polar region are available for comparison [2, 16, 19, 20, 28]. Numerous mathematical formulae for the amplification of perturbations have been derived, e. g. [4, 5], and may be evaluated for their applicability to observed conditions. However, since this paper is essentially descriptive, only a brief explanation is given here and a

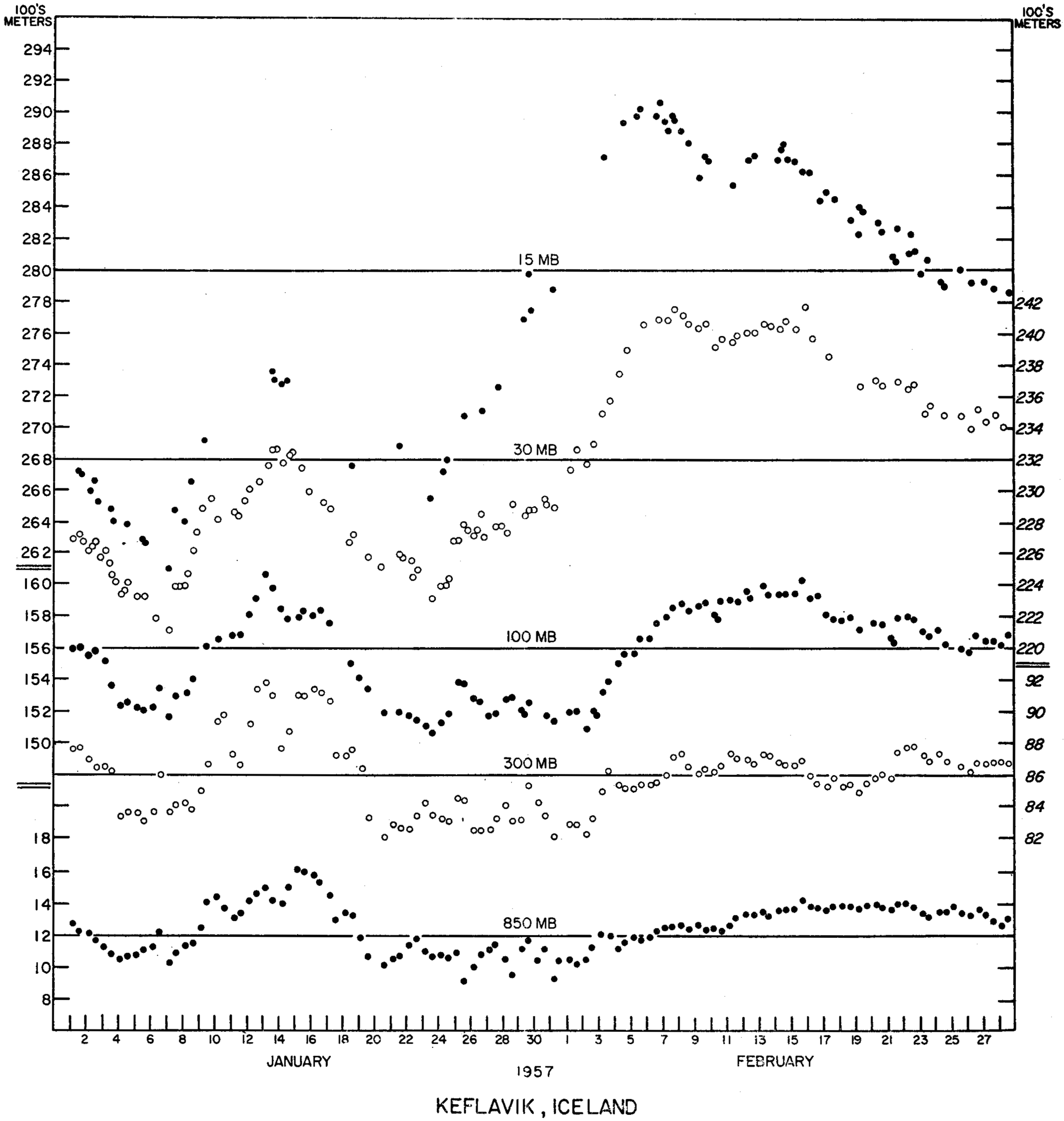

Figure 23.—Height of 850-mb., 300-mb., 100-mb., 30-mb., and 15-mb. surfaces at Keflavik, Iceland, during January and February 1957. Scales for open circle points are on right. Along the abscissa, lines show 0000 GMT for each date.

more complete discussion reserved for a later paper. As one means of limiting the length of this paper interrelationships, similarities, and differences between stratospheric and tropospheric circulations have been discussed only briefly here. A great amount of work needs to be done on the subject, and the attention of research meteorologists is cordially invited.

The Arctic area that is deprived of radiation during polar darkness is perfectly circular and thus we would expect to find a peripheral, horizontal temperature gradient along the circumference. The accompanying vertical wind shear and high level jet stream should be zonal. But the observed jet stream tends to exhibit anticyclonic flow over the Aleutian-Alaskan area and cyclonic flow over eastern Canada and Greenland. From available data, it is assumed that so marked a forced perturbation does not exist in the Antarctic. Thus we have the interesting phenomenon of a topographic feature influencing the circulation of the high stratosphere. Differential radiation from the warm Aleutian ocean area versus that

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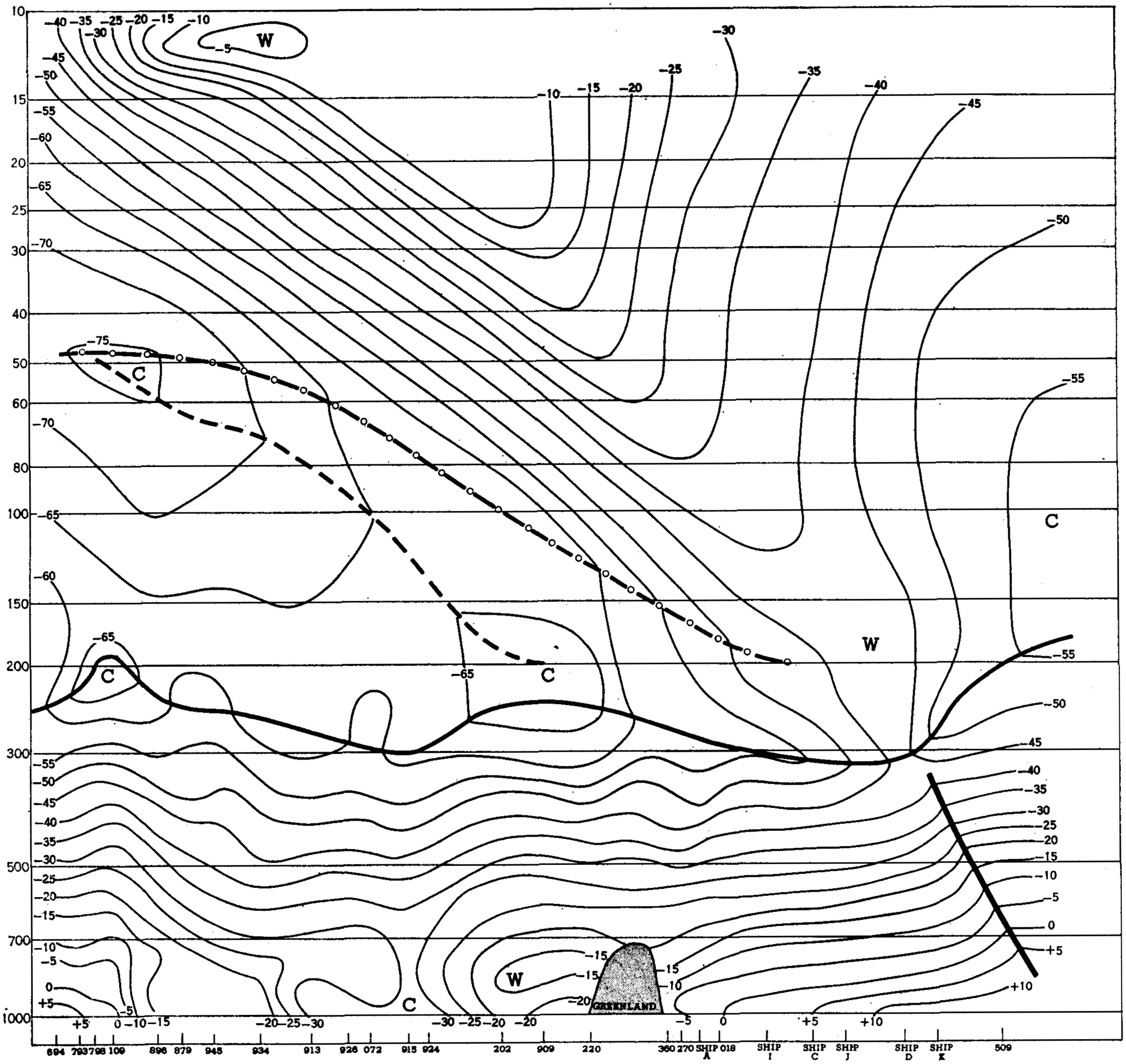

Figure 24.—Cross section for February 6, 1957, from Lajes, Azores (509) to Salem, Oreg. (694), as delineated in figure 22. Very heavy line shows front; heavy line, tropopause; heavy line with open circles, stratopause; and heavy dashed line, lapse rate discontinuity. Isotherms are in °C.

from the cold Greenland ice cap is in the right sense to explain the warmer stratosphere over the Aleutians and thus the presence there of a warm high-level ridge in contrast to the cold stratospheric trough that seems to be normal over western Greenland during the first months of darkness. This same reasoning was used by Schumacher [17] to explain the fact that in the Antarctic, where the surface is colder than in the Arctic, the stratosphere exhibits temperatures 5° or 10° C. colder than have been measured in the Arctic stratosphere. However, the following explanation is given in terms of the vertical and horizontal circulation which may be a contributing, or even the entire, cause of the forced perturbation.

In the high troposphere over Japan flows the most powerful sector of the jet stream, with its position and steadiness determined largely by the horizontal temperature gradient maintained across the Himalaya Mountains [32]. After passing beyond Japan this jet stream decelerates and in so doing forces an indirect circulation in the vertical cross section taken north-south across the central Pacific. In the Aleutian area, the indirect circulation requires rising motion below the level of the jet core and sinking motion above and in consequence forces the formation of a semipermanent cold Low in the troposphere and warm ridge in the stratosphere. On the eastern side of the ridge, the northwesterly flow taps the

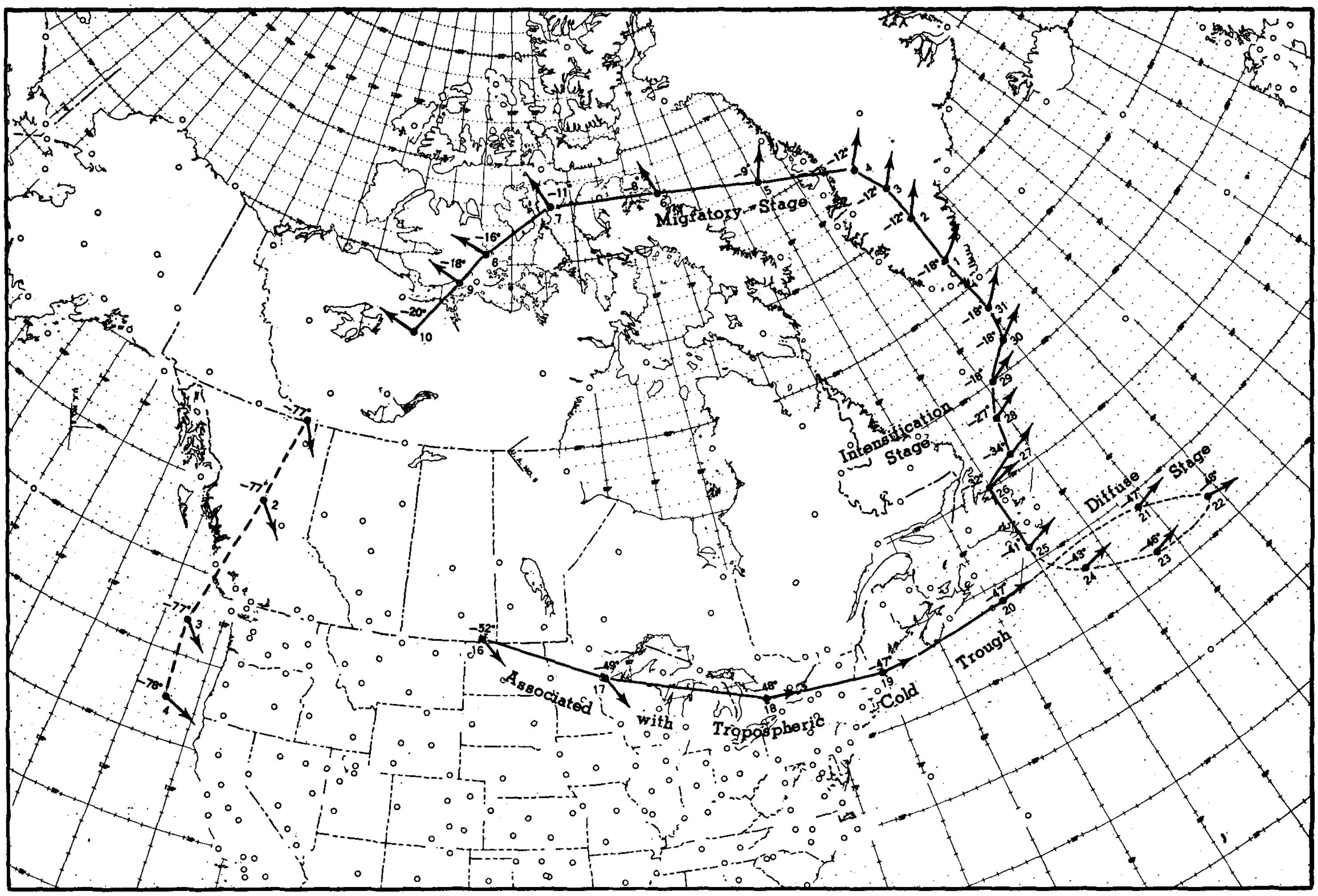

Figure 25.—Trajectory of 25-mb. warm center, January 16 to February 10, 1957, and cold center, February 1 to 4, 1957. Location of centers each day at 1500 GMT is plotted with maximum temperature above, date below, and arrowhead pointing with the wind through the center.

cold air source at the pole, and as the flow, in accordance with the conservation of absolute vorticity, swings around cyclonically near the south of Greenland, the cold stratospheric trough is formed. This explanation seems more convincing than that based on radiation from the surface since warm air not continuously produced by dynamic means could too easily be advected away. That the warm ridge is at times certainly dynamically produced is shown by occasional intensification above cold sources during progression over Canada (1958) or retrogression over Siberia (1957).

Most mathematically derived formulae for amplification of perturbations show baroclinicity or vertical wind shear to be the most critical parameter under the conditions that exist in the upper stratosphere, and baroclinicity approaches the critical value there only on the periphery of the Arctic stratospheric airmass. The Aleutian ridge is a preferred locale for large baroclinicity because it is there that warm air is found nearest the pole. Thus in most years amplification of the Aleutian ridge and its eastward movement away from the area of its origin are important elements in the observed breakdown of the Arctic stratospheric jet stream.

The retrogressive movement of Aleutian ridge and Greenland trough in January 1957 seems to have been an unusual occurrence but was associated with unusually intense activity and so was in accord with the statement by Austin and Krawitz [1] that retrogression is characteristic of stratospheric perturbations during intensification. Possible influences exerted by tropospheric activity were a forced meshing with the anomalous quasi-stationary trough over the western United States and the blocking action of the column of warm air created over the Atlantic by the dissipation of the kinetic energy of the tropospheric jet stream.

The location of the intense 1957 warming on the eastern side of the trough was apparently exceptional but may seem so only because in previous years the eastern portion of the trough was much farther east and so beyond the region of adequate observations. Alternatively, it may eventually be shown that great warming of the eastward portion of the trough is a characteristic of retrogression. In any case, it seems that the exceptional intensity of the warming was at least partially a result of the exceptional strength of the preexisting tropospheric jet stream.

For the principal source of the kinetic energy created during the acceleration and elongation of the stratospheric jet stream and the potential energy represented by the appearance of anomalously warm air in the Newfoundland and Kamchatka areas, we need only to note the simultaneous loss of potential energy by the sinking and warming of the Arctic stratospheric airmass. Still it must be emphasized that the process took several weeks, that friction and eddy motion accounted for a portion of the dispersed energy, and that some of the energy was transformed more than once. The slowness of most stratospheric developments even with vast amounts of available potential energy attests to the influence of strong stabilizing effects, chief of which must be hydrostatic stability.

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